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The fine-scale spatial and temporal variability of hydrologic attributes associated with the process of infiltration in 'nano-catchments' during a rainfall event.

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ABSTRACT

The dynamism of a variety of hydrologic phenomena tied to the process of infiltration are studied here in relation to their spatial and temporal variability within sub-hectare bowl-like depressions, or 'nano-catchments'. The process of infiltration is becoming increasingly important to understand as a result of anthropogenically driven changes to the near-surface soil matrix, which alters this process.

Within the context of infiltration, the spatial variability of soil moisture is assessed under a changing hydrologic regime in south-central Ontario during a rainfall event. With an increase in soil moisture following precipitation events, the spatial auto-correlation increases for both samples that incorporate 15 cm and 30 cm samples. The pattern of soil moisture is influenced by local topographic shape; however this pattern is also altered by the effect of vegetation in the form of active photosynthesizing vegetation and leaf detritus. The effect of vegetation is such that the relationship between topographic gradient and soil moisture is enhanced under active vegetation, while this same relationship is muted under leaf litter.

The variability of infiltration to the point of soil saturation is also assessed. A number of estimates of hydraulic conductivity are used, as well as differing estimates of soil moisture to evaluate the bias of using single point measures versus areal estimates in the modelling of infiltration within these nano-catchments. In conjunction with infiltration modelling, matric potential throughout two nano-catchments is assessed in relation to site characteristics including vegetation, macropores and topographic position.

Conclusions support that in monitoring infiltration and soil moisture cannot be fully represented by single point measurements, even at a sub-hectare scale.

CO-AUTHORSHIP STATEMENT

The following thesis contains material from one manuscript that has been submitted to Hydrological Processes and another that will be submitted to the Journal of Hydrology in the near future.

1. Hydrological Processes

The manuscript titled, "The relative influence of microtopograhy and vegetation cover patterns on fine-scale soil moisture patterns in 'nano-catchments'", is co-authored by P. Andrew-McBride and P.A. Graniero. Field collection and analysis presented was performed by the author. The submitted version appears in Chapter II.

2. Unsubmitted Journal article

The manuscript titled, "On the scaling of infiltration: Spatial and temporal patterns of matric potential and infiltration to saturation in two sub-hectare 'nano-catchments'" is co-authored by P Andrew-McBride and P.A. Graniero. Field collection and analysis was performed by the author. This manuscript appears in Chapter III.

DEDICATION

I would like to dedicate this work to those few pirates who are keeping global warming at bay.

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I would like to thank all those who have aided me in this research. This includes a number of people who have either directly assisted in this undertaking or have provided support along the way.

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Farley too!

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STATEMENT OF ORIGINALITY

I hereby certify that the work embodied in this thesis is the result of original research and has not been submitted for a higher degree to any other University or Institution.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER I
An introduction to the process of infiltration and the dynamism of sub-hectare soil hydrology

1.1 Introduction

Infiltration is the movement of water across the soil surface boundary and into the matrix below. As noted by James and Larson (1976), “infiltration is the key process in predicting or modelling not only hydrologic events, but also in representing water movement and storage for a number of water management practices”. Among the practices reliant on infiltration are agricultural activities and the study of the rate at which groundwater and aquifers are recharged. Of growing importance is the contamination of groundwater, which is directly tied to the downward movement of pollutants via the process of infiltration. With increasing human intervention in this realm it is very important to understand the dynamics that influence infiltration.

This study focuses on elucidating spatio-temporal variability of soil and hydrological characteristics associated with infiltration and the redistribution of water prior to, during, and after a rainfall event. The majority of the data present here within is from a five day period from April 19-23, 2005, additional data was obtained in order to look at the variability of some of the measures over both distance and time. The redistribution of soil water is studied through both changes in the pattern of soil moisture and changes in the hydraulic gradient (i.e. matric suction) throughout the study areas. The study also considers some issues with respect to the accuracy and potential bias in standard estimation techniques at different spatial scales. Each study site serves as a ‘nano-catchment’, or sub-hectare bowl-like hydrological unit, which is assumed to be bounded and hydrologically isolated at the near surface by local plateaus or ridges. The term was chosen to reflect that size and scale of the features under study are smaller than that what is conventionally termed a ‘micro-catchment’ by hydrologists. Through a variety of temporally and spatially intensive field measurements the dynamism of hydrologic gradients and patterns are studied in order to relate these phenomena to local attributes of the nano-catchment. The vertical and lateral movement of water through the near-surface soil matrix is studied using both field and modelling techniques. The pattern of soil moisture at each site is assessed with respect to the local microtopography and vegetation patterns. Both active vegetation and leaf litter are studied in order to associate the pattern of soil moisture to these variables. Matric suction is also monitored in terms of these two variables at specific points within each nano-catchment. The influence of macropores on the hydrologic processes within the nano-catchments is also considered. These assessments are then tied to the variation of infiltration to saturation, infiltrometer tests and estimates of hydraulic conductivity across each bowl. Bias is discussed in terms of using single point measurements to assess variability and areal totals of infiltration.

Our approach of tracking the hydrologic and hydraulic characteristics of the sub-hectare nano-catchments, where the scale of interest ranges from 1750 to 6233m^2, separates this study from previous field studies conducted at point scale ($10^{-1}m^2$) (Hansen et al., 1999; Hansen, 2000), plot scale ($10m^2$) (Paige and Stone 2003), or regional scales ($\sim 10^8 m^2$) (Sullivan et al., 1996; Zhang et al., 2000; Scozzafava and Tallini, 2001). Typically, field studies of infiltration variability have considered the influence of soil (Al-Turbak, 1996) or vegetation characteristics (Stothoff et al., 1999; Bharati et al., 2002). While there has been considerable research focusing on the variability of the parameters, little research has been disseminated on the basis of microtopography and the relation of the attributes noted above to hydraulic conductivity and matric potential, which all influence infiltration dynamics. Soil moisture has been studied fairly extensively in terms of the relative role of topography, however the debate is still continuing with respect to its exact role (Devito et al., 2005; Western et al., 2005; Tromp van Meeveld and McDonnell, 2005). While Fox et al., (1998a, 1998b) looked at movement of moisture in a microtopographic laboratory environment, and Grayson, Western and associates (Grayson and Western 1998, 2001; Western et al., 1999) have looked at soil moisture in relation to meso-scale topography on the catchment scale, a gap still exists in our understanding. The study presented here provides a glimpse into the variability of these hydrologic attributes at this intermediate scale, while also differing in terms of the rapidity of the field collection methodology used to capture both terrain and hydrologic attributes over the two sites of interest.

In previous studies the explicit characterization of micro-terrain at scales larger than plot scales (10^1 m^2) has been difficult. Here however, through measurement technology developed in parallel with this project, micro-terrain and fine-scale point measures of soil moisture are more easily collected and therefore the variability of these features at this scale are characterized in a manner that differs from those of earlier studies. The use of the Multi-purpose Environmental Modelling Facility (MEMF) Lab's ProbeFusion software (Graniero and Miller, 2003) allows for the temporal and spatial variability to be studied in a manner that cannot be done using standard methodology. The data density encountered in this project ranges from a few points over the nano-catchments for conventional soil analysis, infiltrometer tests and matric suction to hundreds of points per hectare for daily soil moisture and thousands of points per hectare for elevation. While moisture content and matric potential are generally monitored concurrently, the mobility of the equipment in this study again differentiates it from previous studies. Moisture content is generally observed at a single location or a group of stationary locations over a field site; in this study moisture measurements are made at various temporally dynamic points. While Grayson and Western (1998) used a dense collection of stationary monitoring locations, they focused on a temporal scale of one year to investigate long-term variance, whereas this study focuses on short-term temporal variability on the scale of minutes to days and on spatial variability on the scale of meters to tens of meters.

The results of this study will be forwarded to the Canadian Center for Remote Sensing to aid in determining whether their current meteorological stations provide an accurate representation of the soil moisture flux over an area, or just localized point information. These types of assessments are critical for improving their parameterization methods for their regional hydrological and aquifer recharge modelling efforts on the Waterloo and Oak Ridges aquifers.

1.2 Background Terminology

The terms associated directly with infiltration include the infiltration rate of the soil i [L/T], and the infiltration capacity or infiltrability I [L/T]. The infiltration rate of the soil is the depth-equivalent of water that penetrates the soil from the surface per unit time. The maximum rate of infiltration is termed the infiltration capacity or infiltrability. If precipitation surpasses the infiltration capacity surface saturation occurs, which results in either ponding or overland flow. The capacity is a result of the type of soil and the antecedent moisture content, which is largely an effect of prior precipitation history and site-specific parameters of the soil. The physical characteristics of the soil matrix that affect infiltration are numerous, but the primary characteristics are the porosity ϕ [L^3/L^3], bulk density ρ_b [M/L^3], organic matter [L^3/L^3], and texture. These attributes are tied to the hydrologic phenomena that alter the rate and capacity of infiltration at a given site. These measures can also be used in estimating the hydraulic attributes of the soil unit under study. Termed pedotransfer functions (Bouma, 1989), these equations represent the relative influence of the noted parameters on the hydraulic attributes of a soil through multiple linear regression analysis. The hydrologic factors associated with the rate of infiltration and infiltration to saturation, which is the amount of water infiltrating the ground surface prior to surficial saturation, are also numerous, but are dominated by the matric suction ψ [L], the soil moisture content θ, and the hydraulic conductivity K [L/T] within the soil.

The term 'matric suction' ψ [L], is interchangeable with matric potential and tension head, all of which are used in the literature. In essence, ψ is the negative pressure at which water is held in the matrix of an unsaturated soil (Dingman, 2002). The matric potential, in conjunction with the gravitational potential (typically measured by the elevation above an arbitrary datum), is termed the hydraulic head h [L], and represents the total potential energy at a point. The gradient in potential energy between points determines the movement of water within the matrix (Jury et al.,1991).

A number of factors contribute to the pressure exerted on water at a given location. Among these are the overlying weight of the soil matrix and water, the force of gravity, and the effect of solutes, which can be assessed individually (Tindall and Kunkel, 1999). Additionally, the water content of a given soil can be represented in two ways. The first approach is the gravimetric moisture content θ_g [M/M], which is simply the ratio of the mass of water to the mass of the soil containing the water. The second and more common approach in monitoring soil moisture content is the volumetric method θ $[L^3/L^3]$, which is the ratio between the volume of water and the volume of the soil (including pore spaces) containing the water. In the study undertaken here all soil moisture is assessed volumetrically.

The hydraulic conductivity K [L/T] of a soil is the general rate at which water flows through a porous medium, in this case the soil profile. In unsaturated conditions, the hydraulic conductivity, $K(\theta)$ or $K(\psi)$, is a function of the soil moisture or the matric potential, respectively. As the soil moisture increases, $K(\theta)$ also increases and reaches a maximum when the soil is saturated. At this point the hydraulic conductivity is denoted K_s. The relationship between ψ and θ can be defined as the soil moisture capacity function, also termed the soil water retention curve or the soil water characteristic curve, which is a non-linear sinusoidal relationship. This relationship differs during wetting and drying as a result of hysteretic processes. During drainage it takes a larger amount of pressure to dewater small pore spaces and hence the larger pores above these pores remain full of water. Alternatively, during wetting, these small pores are filled relatively easily, and the large pores allow for the flow of water through them into small pores. Another contributing factor to this effect is the presence of air in the pores which cannot be removed no matter the pressure exerted during wetting and the change in the curvature of the meniscus of pore water during filling and drainage (Stephens, 1996; Tindall and Kunkel, 1999). As a result of hysteresis, at a given pressure there will be a higher moisture content during drainage than during wetting. In the typical approach to modelling the characteristic curve via pedotransfer functions the effect of hysterisis is neglected (van Genuchten, 1980; Rawls et al., 1983; Saxton et al., 1986).

There are several terms associated with the relationship between soil moisture and matric potential. The bubbling pressure ψ_b [L], for instance is the tension at which the soil moves from a saturated state to an unsaturated state. This is important because there is a range of low pressures (i.e. suction is close to zero) which will allow the soil to remain saturated. This point also serves as the inflection point in the soil characteristic curve. The pore size distribution index λ is a dimensionless, indirect measure of the connectedness of pores in the soil matrix. Both ψ_b and λ affect the wetting front suction, ψ_{wf} [L], which is the suction applied by the 'dry' soil to the advancing wetting front. In reality the wetting front is often not a sharp frontal movement through the soil, however it is this assumption that allows for the modelling of infiltration using most models derived from Green and Ampt (1911). The two variables can be estimated through the use of pedotransfer functions (Rawls et al., 1983; Saxton et al., 1986).

1.3 Processes affecting the distribution of water in the soil matrix

The interactions of the above parameters influence the mechanics that control the amount and rate of water flowing through the medium. Darcy's Law (1856) describes q_x [L/T], the general movement of water through a homogeneous medium:

$$q_x = K(\theta)\frac{dh}{dx} \tag{1.1a}$$

or

$$q_x = K_s\frac{dh}{dx} \tag{1.1b}$$

where K_s is the saturated hydraulic conductivity, $K(\theta)$ is the unsaturated conductivity and dh/dx is the hydraulic gradient. h is the total energy potential comprised of gravitational and capillary potential (Smith, 2002). This general equation can also be used in estimating horizontal flows with the obvious removal of the effect of gravity. It is Darcy's Law which is the foundation for much of the research conducted on the infiltration process.

Green and Ampt (1911) provide an example which extends Darcy's description into one of the fundamental equations of infiltration. The Green-Ampt model is "based on the idea that infiltration can be depicted by a very steep wetting front behind which the water content has a constant value θ" (Aoda et al., 1988). The wetting front is the depth to which the water from a particular input event has moved downward, generally symbolized L [L].

Green and Ampt describe infiltration i [L/T] as:

$$i = K_s \left[\frac{(d_p + L) - \psi_{wf}}{L} \right] \tag{1.2}$$

where d_p [L] is the depth of ponding at the surface, K_s is the hydraulic conductivity behind the wetting front and ψ_{wf} is the pressure at the wetting front. Behind the wetting front the soil is assumed to be saturated, however this is not generally the case for the entire profile behind the front, and thus $0.5K_s$ is sometimes used (McCuen et al., 1981; Vieux, 2004). However, this is neglected in Chapter III in order to account for the offsetting effect of macropores on this process.

Green and Ampt was extended by Mein and Larson (1973) in order to describe a two-stage infiltration event. While the Green-Ampt description of infiltration occurs only after ponding at the surface has occurred, Mein and Larson developed the following equation to quantify the amount of infiltration and the time of ponding:

$$F_s = \frac{\psi_{wf} \cdot \Delta\theta}{r / K_s - 1} \tag{1.3}$$

where F_s [L] is the cumulative infiltration up until ponding, $\Delta\theta$ [L/L] is the soil moisture deficit (θ_s-θ_i), θ_s is the saturated soil moisture content and θ_i is the initial soil moisture content, r [L/T] is the intensity of the precipitation input and ψ_{wf} is the pressure at the wetting front. This equation is fairly intuitive in the fact that when $r > K_s$ and when $\theta_s = \theta_i$, ponding will occur. While Mein and Larson altered the Green and Ampt approach, so have a number of others to account for heterogeneities. For example, Flerchinger et al., (1988) used a model to estimate infiltration in a layered soil, where the rate of infiltration varies primarily as result of the hydraulic conductivity of the individual soil layer.

1.4 Field Measurements

There are a number of methods used to measure infiltration, or the parameters used to estimate it, in the field. Tools range from manual monitoring to electronic or nuclear probes. The goal of the study and the field setting must be considered when selecting the appropriate tool.

Infiltrometers, as the name implies, monitor infiltration into the soil and can be deployed fairly easily (Sanders, 1998). Infiltrometers are basically a ring or rings into the soil surface in which water can be added. The depth of water is monitored in order to determine the rate at which it percolates downward through the matrix:

$$i(t) = \frac{W - \Delta H \cdot A}{\Delta t} \tag{1.4}$$

where $i(t)$ [L/T] is the infiltration rate at a time t, W [L^3] is the volume of water added to the ring during the time period Δt, ΔH [L]is the change in the depth of ponded water and A [L^2] is the area of the infiltrometer ring.

There are also a number of indirect methods for monitoring soil moisture. One such method is Time Domain Reflectrometry (TDR) which consists of two prongs of 30 cm or less that serve as waveguides for pulses of electromagnetic radiation (Jury et al., 1991). The time that it takes for the pulse to return to the source of the radiation is measured to determine the permittivity or dielectric number, P. The dielectric number is a measure of the conductance and capacitance of a medium:

$$P = \frac{tc}{L_w} \tag{1.5}$$

where L_w [L] is the length of the waveguides, t [T] is the transmission time and c [L/T] is the speed of light. The calculation to determine the θ from a given permittivity was estimated by Topp et al., (1980) as:

$$\theta = -5.3\times 10^{-2} + 2.92\times 10^{-2} P - 5.5\times 10\times 10^{-4} P^2 + 4.3\times 10^{-6} P^3 \tag{1.6}$$

where P again is the permittivity and θ is the volumetric moisture content. This equation varies with differing TDR units, and most manufacturers supply calibration information.

A number of other in-situ sensors are used in the monitoring of volumetric soil moisture content, ranging from electrical resistance blocks generally constructed from gypsum, to neutron moisture meters and gamma-ray scanners. Remote monitoring is also possible through the use of microwave sensors, such as RADARSAT (Huisman et al., 2001). Microwave sensors are again influenced by the soil's dielectric properties in terms of the scattering and hence the strength of the returned signal. Remote satellite sensors have the advantage of being able to monitor soil moisture over very large extents however, as with all satellite sensors, the return time over which monitoring can be done is long (Yoo, 2001) and calibration to ground data can be difficult.

The use of tensiometers to measure the pressure exerted through matric potential is well documented (Hornberger et al., 1998; Zhang et al., 2000; Weiler and Naef, 2003). Tensiometers measure the potential across a ceramic membrane placed in the soil. When the tensiometers are placed in the soil the water is held at atmospheric pressure, thus the liquid flows across the membrane as a result of the regional tension. The depth at which the tensiometers are placed into the soil allows for vertical gradients to be established, while lateral gradients can be evaluated by placing tensiometers over an area.

1.5 Sites of Study

The sites examined in this study are significant as a result of their geologic and socio-economic position in Southern Ontario. The study sites lie around the Oak Ridges Moraine (ORM), which is of special interest because it serves as a major area for groundwater recharge in the Metropolitan Toronto Region. The ORM stretches from approximately Rice Lake at its eastern margin to the Niagara Escarpment to the west (Fig.1.1). This glacial feature has been the focus of a great deal of study, specifically as to its origins (Barnett et al., 1998).

Two primary locations around the Moraine were investigated, with one site situated on the ORM and another on the Niagara Escarpment. The locations chosen were deemed suitable as a result of a number of characteristics including the ease of access, the local setting in regards to both hydrology and topography, and local vegetation characteristics. Glen Haffey (Fig 1.2 and 1.3) is located in UTM zone 17N at coordinates

N: 4865882; E: 584560, while Crawford Lake Conservation Area (Fig 1.4) is located in N: 4813460; E: 585377. These sites are hydrologically important as noted above, because of their location around the ORM. Geomorphologically they are significant because of the differences in sedimentology among the study sites. The Glen Haffey site lies on glacial till, specifically Halton Till which is a superfluous deposit across the ORM (Barnett et al., 1998). The Crawford Lake site straddles Wentworth Till and an area of exposed bedrock. This site is located on the Waterloo aquifer which also provides potable water and water used in agricultural processes.

While the sites are of physical importance for a variety of reasons, the significance of the sites is amplified by the considerable human modification of the surrounding environment. Due to the proximity of the ORM and the Waterloo aquifer to Metro Toronto they are being increasingly altered as a result of human activity. The surface overlying these aquifer recharge zones is undergoing change in a variety of ways. The dramatic population growth in the region has led to increased surface sealing over the area and thus a change in the local hydrologic budget. This 'sealing' is a result of the construction of road and sanitary networks that are designed to efficiently and rapidly remove water from the area. These networks have altered the movement of water within the soil, and have also increased the rate at which input events reach the streams and rivers in the area.

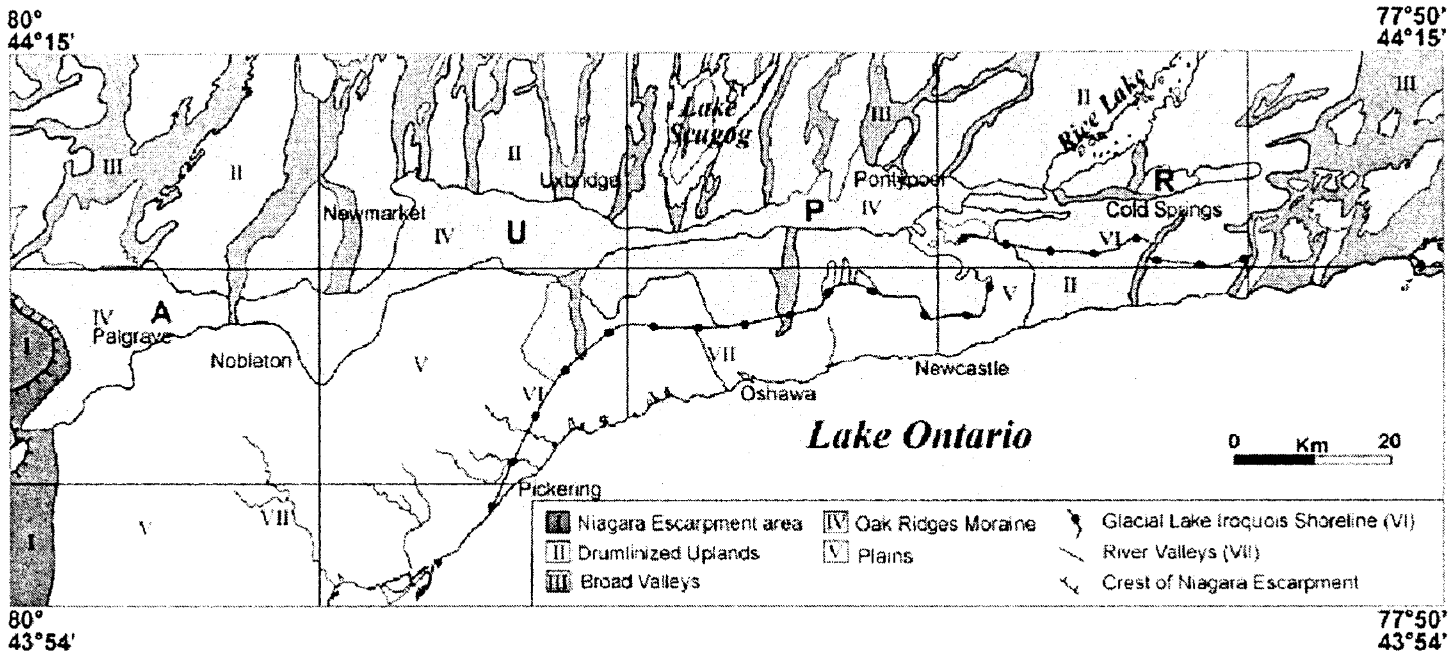

Figure 1.1 Oak Ridges Moraine (Barnett et al., 1998)

Figure 1.2 Glen Haffey Field Site – Western Margin

Figure 1.3 Glen Haffey Field Site – Eastern Margin

Figure 1.4 Crawford Lake Field Site

1.6 Analysis and Modelling

A variety of tasks were undertaken in order to evaluate the data itself and the processes which lead to variability in soil moisture, matric suction and infiltration throughout both sites. Where point measurements were made, estimates regarding the rate of infiltration were assessed. The Green-Ampt (1911) model serves as the basic tool in providing point estimates of the infiltration rate, and Mein and Larson (1973) serves as a tool to evaluate infiltration prior to saturation at differing spatial resolutions. Green-Ampt is used as a baseline in this study because all of the parameters are easily discernable, which cannot be said for a number of other models such as the Soil Conservation Service's Curve Number SCS-CN (1972).

The variability in the capacity and rate of infiltration has been studied at a number of scales from the watershed down to plot size areas. It is fairly well understood that the hydraulic components of the process vary significantly over relatively small distances (Springer and Lundy, 1987). Overall, the hydraulic conductivity of a soil as a rule varies more than either soil moisture content or matric potential (Jury et al., 1991). Grayson and Western (1998) studied the spatial variability of θ in an attempt to ascertain the variability of soil moisture over field sizes from 0.10 to 27 km^2. Paige and Stone (2003) attempted to establish the spatial variability of infiltration at a much smaller plot scale ($12m^2$). Here, spatial analyses of soil moisture are undertaken using variography which assesses the spatial auto-correlation across a region at a point in time (Western et al., 2004). Greater detail on past studies regarding soil moisture variability is presented in Chapter II.

Areal estimates of infiltration over both bowls are also calculated using a variety of techniques as discussed in Chapter III, however in order to evaluate this process areally a number of analyses were undertaken. The most fundamental of these is the construction of a very fine-scale digital elevation model (DEM) which serves as the basis for much of the research. The DEMs are constructed using standard Gaussian variograms and kriging procedures as described by a number of authors (Jury et al., 1991; Ersahin, 2001; Lo and Yeung, 2002). These DEMs provide the platform for the analysis of the pattern of soil moisture and matric suction with reference to topographic indices. A number of topographic indices are studied, with attention to the commonly used Wetness Index (Beven and Kirkby, 1979). It is generally assumed that topography is the driving mechanism in the subsurface redistribution of infiltrating waters, however as noted throughout this study the influence of topography is not explicit at all scales and times. These indices allow for the assessment of the role of topography or the lack thereof. This relative role is then compared to other nano-catchment attributes such as vegetation and the presence of macropores in attempting to explain the variability of soil moisture, matric potential and hydraulic conductivity.

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CHAPTER II
The relative influence of microtopography and vegetation cover patterns on fine-scale soil moisture patterns in 'nano-catchments' during a rainfall event

This chapter has been submitted to Hydrological Processes as a manuscript titled, "The relative influence of microtopograhy and vegetation cover patterns on fine-scale soil moisture patterns in 'nano-catchments'", it is co-authored by P. Andrew-McBride and P.A. Graniero.

2.1 Introduction

The state of soil moisture in the near surface affects the amount of runoff that is generated during a precipitation event and to a great extent controls the amount of water that infiltrates into the soil (Philips, 1957). This in turn plays a role in recharging groundwater systems. The amount and spatial variability of soil moisture also has a significant influence on the amount of evaporation that occurs from a soil, and thus serves as a major control in partitioning sensible and latent energy (Albertson and Montaldo, 2003; Illston et al., 2004). This division of energy not only controls the local energy budget, but at broader scales can have an aggregate effect on the global energy budget and therefore the climate. In addition to meteorological and climatological response, understanding soil moisture is critical to managing anthropogenic activities, including decisions on farming, urbanization, and contaminant control (Chen et al., 2004).

Soil moisture variability is scale-dependent, since a number of different processes alter the pattern at any given location (Gomez-Plaza et al., 2000; Blöschl, 2001; Lookingbill and Urban, 2004; Petrone et al., 2004). Studies have examined the scale-based effect of climatological and meteorological controls (Carrey and Woo, 1999), bedrock properties and proximity to the surface, soil properties and vegetation controls (Albertson and Kiely, 2001; Hupet and Vanclooster, 2002; English et al., 2005). The spatial variability of these characteristics plays a major role in the pattern of soil moisture from multiple perspectives. For example, Pariente (2002) points out that the presence of a shrub can alter the soil moisture content at a location by providing shade, thereby decreasing evaporation and therefore soil moisture loss, but also by the presence of roots which as a result of plant transpiration removes water from the soil. Thus, it is difficult to account for the aggregate variability soil moisture arising from all of these controls.

While vegetation, soil properties and meteorological differences have been studied, most research to date has focused on the pattern of soil moisture in relation to topographic position. The use of a wetness index (Beven and Kirkby, 1979) has become a standard practice in monitoring soil moisture patterns and the distribution of water in relation to topographic position. This index provides a proxy for the relative moisture at a point in terms of the local slope and the upslope catchment area (Grayson and Western, 2001; Green and Erskine, 2004; Ibbitt and Woods, 2004; Western et al., 2004). Areas of convergence in a landscape and specifically those sites that have the potential to receive water from large upslope areas will be comparatively moister than those on ridges or at higher relative elevations. However, the effect of topography on soil moisture and other hydrologic processes, especially during dry periods below field capacity, is of great debate (Devito et al., 2005; Tromp van Meerveld and McDonell, 2005; Western et al., 2005). Typically the significance of topography can be considered as a result of several other variables including the proximity of bedrock or some impervious layer, the relation between precipitation and evaporation at a site, and a number of other micro-meteorological controls. It has also been generally assumed that during periods of high soil moisture at or near saturation, topography has a far greater significance (Ridolfi et al., 2003). Even this assumption has been challenged lately (Wilson et al., 2005) and the role of topography at saturation cannot be assumed to be the same between sites. While these studies have focused on the catchment scale, it is also important to understand the potential role topography has on both the temporal and spatial variability at a much finer scale.

For typical programs that monitor catchment or aquifer hydrology, sparsely distributed networks of sensors are situated to monitor a variety of hydrological and micrometeorological variables (Maloley, 2004). These sensor stations are generally placed in areas where measurements will be least affected by confounding elements. However, because of the limited spatial density of these sites and their typical placement (i.e. open fields with little or no slope) much of the variability that arises from topographic change, vegetation differences, etc. are not considered. This is especially important because, as noted earlier, these are the elements which influence soil moisture and the associated pattern.

The primary purpose of this paper is to determine at very fine temporal and spatial scales, the relative roles of topography and vegetation cover in the control of soil moisture variability at a point and in reference to the overall distribution of water in the near surface soil matrix over a very small region. We attempt to examine the pattern of soil moisture in both time and space in what we term a 'nano-catchment'. We define nano-catchment as a bowl-like depression that is approximately a hectare or less in size, acting as an isolated catchment with respect to surface and near-surface flow. This term allows for a bridge between the micro-catchment (10^5 -10^7 m^2) scale and the plot ($10^1 m^2$) scale study unit. Within this nano-catchment it is assumed that considerable spatial variability will be found and that this spatial pattern will change through time as overall soil moisture changes (Grayson and Western, 1998; Ridolfi et al., 2003; Shcume et al., 2003; Ibbit and Woods, 2004; Lookingbill and Urban, 2004; Western et al., 2004).

We concentrate on a short time scale (hours to days) which provides a basis for monitoring the temporal stability of soil moisture at and within the study area (Grayson and Western, 1998). It also permits a comparison to be made of the spatial patterns at higher and lower soil moisture levels over short periods of time, thereby allowing several factors to be assessed with respect to soil moisture pattern, including topographic gradient, macropores, the proximity of bedrock, and differences in vegetation cover (both growing vegetation and leaf litter).

From a technological standpoint, major advances are being made in collecting finer topographic detail via LIDAR and in collecting finer land cover detail via high resolution satellite imagery. Therefore our secondary aim is to evaluate whether greater effort in gathering fine-scale topographic data sets or fine-scale imagery would help to more accurately assess the spatial pattern of soil moisture, and subsequently infiltration and groundwater recharge.

2.1.1 Measuring Soil Moisture

Point measurements of soil moisture content are made using a number of different methods, including gravimetric procedures, time domain reflectrometers (TDR) and neutron probes. While gravimetric assessment allows for soil moisture content to be assessed very accurately it is prohibitively time consuming in terms of both sample collection and analysis. Neutron probing can be quite accurate, but is limited in achieving a dense spatial sample and poses some radiation risk. TDR estimation is based on the dielectric properties of water in relation to soil and air, with Topp et al., (1980) pioneering the method and others following with local calibrations (Carey and Woo, 1999; Grayson and Western, 2001; Vaz et al., 2002). These studies have solidified the use of TDR as the dominant tool for determining soil moisture at a point in space. However, determining the soil moisture condition on an areal basis has not been quite as straightforward.

The use of a variety of geostatistical tools have been employed to extend point measures of soil moisture to represent areal patterns. The most widely used measure to describe the spatial relationship among points of known soil moisture is the variogram (Feng et al., 2004; Petrone et al., 2004; Western et al., 2004) and the subsequent generation of an areal measure using a variety of kriging procedures (Bardóssy and Lehmann, 1998; Jost et al., 2005). Anisotropy has been studied (Schume et al., 2003), however onmidirectional variograms have been deemed sufficient (Western et al., 1998b; Western et al., 2004). One issue in all types of kriging procedures is the hydraulic connectivity that presents itself in the pattern of soil moisture. While topography is not always an explicit control in certain areas, gullies are typically prone to having higher soil moisture contents for example and representing this temporally varying connectivity is difficult (Grayson and Western, 2001).

In addition to using geostatistics a number of areal based measures have been explored with varying degrees of success. The use of remotely sensed imagery has been an area of significant focus over the last two decades. Most of these investigations focus on the use of the microwave portion of the electromagnetic spectrum (Mohanty and Skaggs, 2001; Hoffman, 2005). The size of the nano-catchments under study in this present paper are approximately the size of one or two pixels (25 * 28 m) in a satellite-based synthetic aperture radar systems including RADARSAT-1 (Alvarez-Mozos et al., 2005). This study therefore provides additional insight into the current research that has been examining the intra-pixel soil moisture variability (Mohanty and Skaggs, 2001). While variability is assumed to exist, understanding the degree of this variability is desirable. Remotely sensed soil moisture indices are somewhat limited because microwaves rarely penetrate the surface to a depth greater than 5 cm (Mathieu et al., 2003). The controls within a heterogeneous landscape must then also be acknowledged in an attempt to estimate soil moisture in the near surface to a typical rooting depth of 30 cm. Another major problem to date is the presence of plant canopies, which prevents direct sensing of the ground surface. It is therefore key to understand the inherent variability that exists within these spatial units in addition to other microsite controls such as soil texture, macropores of faunal and floral origin, and bedrock depth (Harden and Scruggs, 2003).

Recent research has looked at extending our understanding of electrical permittivity in relation to soil moisture content, moving from TDR to the use of ground penetrating radar (GPR) to provide an areal estimate of soil moisture at the field scale (Huisman et al., 2001; Grote et al., 2003; Huisman et al., 2003; Lunt et al., 2005). This work has had some degree of success in monitoring the pattern of soil moisture at the field scale, however a number of issues have been encountered which must be considered. As Lunt et al., (2005) claim, the use of GPR is limited because the depth of reflectors must be established in order to properly calibrate a transect for the analysis of soil moisture. Each field site therefore will need to have an in depth subsurface survey undertaken in order to obtain a soil moisture field.

For these reasons Wilson et al., (2005) note that "no method of reliably measuring the spatial distribution of soil moisture content in the root zone yet exists" (p.43).

Therefore, it appears that point-based measures will continue their prevalence until direct areal based measures become more readily available and accurate. Thus we use TDR for point measurements and geostatical derivations provide areal estimates of the soil moisture distribution over the respective study sites.

2.2 Study Sites

The two study areas are located in south-central Ontario, Canada within the Glen Haffey Conservation Area (43°56'28"N; 79°56'47"W; 416 m asl) and the Crawford Lake Conservation Area (43°28'29"N; 79°57'10"W; 306 m asl) (Fig. 2.1). The first site is situated on the Oak Ridges Moraine (ORM), while the latter is located on the Waterloo Aquifer, both of which serve as sites of major groundwater recharge for aquifers which are under increasing anthropogenic pressure (Sharpe et al., 1996). The Glen Haffey region also serves as the headwaters for the Humber River terminating at Lake Ontario, one of the larger river systems of the area.

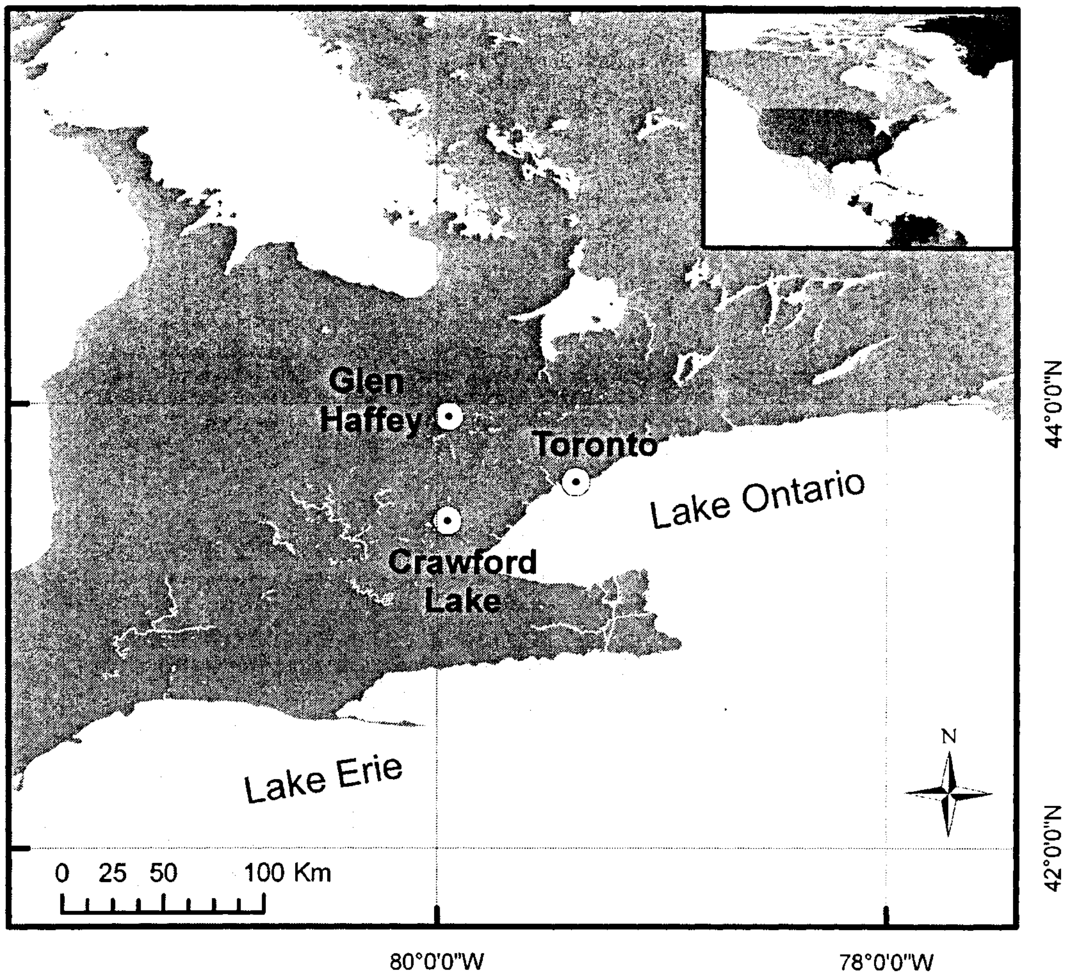

Figure 2.1 Locations of field sites in southern Ontario, Canada

2.2.1. Glen Haffey

The Glen Haffey site is composed of glacially deposited till resulting from the proximal actions of the Laurentide Ice Sheet (12-13 ka B.P.) during retreat with the actual processes by which the ORM formed still under debate (Barnett et al., 1998). The site is composed of hummocky terrain, which is typical of the Palgrave subunit of the ORM (Barnett et al., 1998) and makes it ideal for this study. The specific depression (nano-catchment) chosen for study is 1766 m^2 in size with a relief of approximately 4 m (Fig. 2.2 a). The A horizon is roughly 30 cm in depth and underlain by a cobbly till deposit. The composition of the A horizon varies throughout the site, with ranges of 7-30% gravel, 53-82% sand, and 10-28% fines. While the soil texture is generally sandy throughout the site, the textural variation cannot be wholly neglected with respect to the resulting soil moisture pattern. This nano-catchment has significant topographic variation (mean slope 7.5%, maximum slope 19%). This variation in slope is a result of glacio-fluvial deposition, as mentioned. There are also several erratics at the surface.

The vegetation of the area is composed of mixture of tall fescue (*Lolium arundinaceum*) and Kentucky bluegrass (*Poa pratensis* L.), along with white clover (*Trifolium repen* L.) and small areas of sphagnum (*Sphagnum ssp.* L.). The vegetation is roughly uniform throughout the growing period as it is maintained by the conservation authority. The eastern margin of the study area is flanked by a deciduous forest stand approximately five meters from the bowl, which provides shade during mid- to late-afternoon and also provides a source of leaf litter during the fall which is still present throughout the early part of spring. The leaf litter is spatially variable, with the portion of the nano-catchment proximal to the stand of trees having a greater O_i horizon.

2.2.2 Crawford Lake

The Crawford Lake site is located 50 km south of the Glen Haffey site and is underlain by the Niagara Escarpment. This geologic feature controls much of the surficial topography in the region and served as a barrier in the proximal formation of the ORM (Barnett et al., 1998). Overall, surficial sediments are fairly thin, with limestone bedrock outcrops prevalent throughout the region surrounding Crawford Lake. The area around Crawford Lake has served agricultural purposes for much of the recent past, however the region is increasingly being urbanized.

The specific nano-catchment is 6233 m^2 in size, with a relief of roughly 3.5 m (Fig. 2.2 b). The topographic variation is also significant (mean slope 4%, maximum slope: 23%). Unlike the Glen Haffey nano-catchment this site is not entirely bowl-like, in that the gradient continues beyond the study site and thus would serve as a surficial outlet. The sand content of the four soil samples taken at Crawford Lake ranged between 50% and 62%, while containing 10-21% fines and 14-22% gravel. The samples again are generally sandy, but the texture does vary horizontally between in the four samples taken at 15 cm depth throughout the site.

The vegetation consists of two zones, with a narrow transitional area that serves as a boundary. The western section of the site consists of a fallow agricultural zone, which until recently was used for corn (*Zea Mays* L.). However this area is currently in the very early process of being returned to native vegetation and is still relatively bare. The eastern margin of the site consists of successional vegetation where tall grasses (*Lolium arundinaceum*) predominate with other native plants mixed in. The eastern boundary of the study area is flanked by a deciduous forest as well as a coniferous plantation.

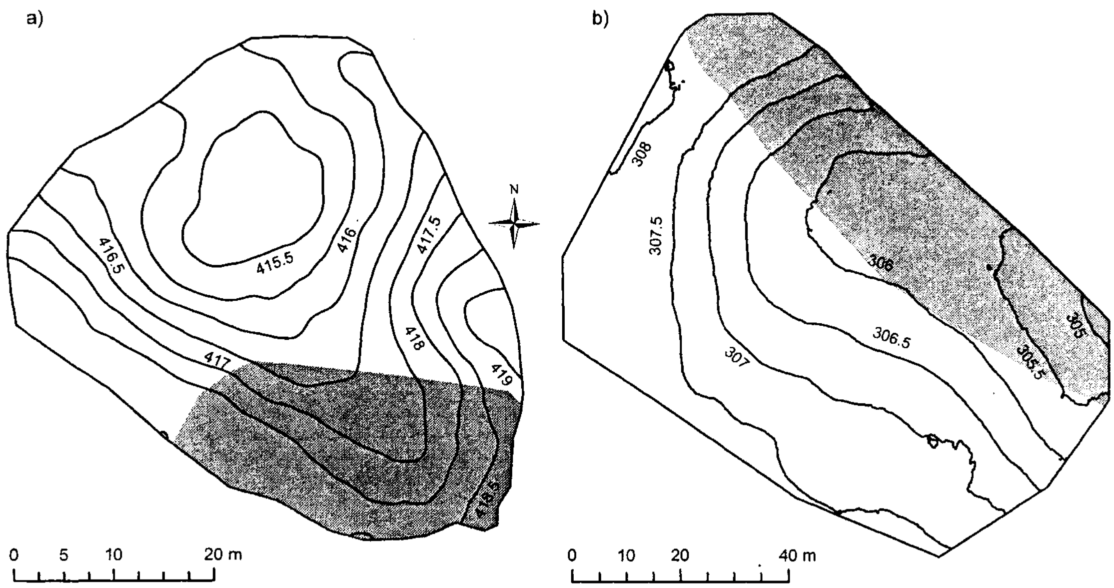

Figure 2.2 The two nano-catchments that serve as the foci of the study: a) Glen Haffey and b) Crawford Lake. The shadowed areas represent leaf litter and vegetated areas.

2.2.3 Climate

The precipitation in the area around both sites peaks during the summer months with the mean monthly precipitation reaching a maximum of 95.6 mm in August and a minimum of 50.9 mm in February, with a mean annual precipitation of 891 mm (Environment Canada, 2004). The Glen Haffey site in particular receives large amounts of convective precipitation throughout the year as a result of the prevailing wind patterns that travel from Lake Huron during ice-free months. This leads to very spatially confined storm events that can deposit large amounts of precipitation over short periods of time. Evaporation at both sites exceeds precipitation for a large part of the year with August having a mean monthly evaporation of 117.3 mm (Environment Canada, 2005). Evaporation is negligible during the winter months, whereas precipitation exceeds evaporation during the fall.

The Glen Haffey site is in close proximity (less than 2 km) to a meteorological station operated by Natural Resources Canada (Maloley, 2004). This meteorological station monitors soil moisture at four depths in the soil and provides a reference for our data collection. The station also provides data regarding precipitation throughout the year. Although the Crawford Lake site is situated at a significant distance from this weather station, other site options were no closer to a station.

2.3 Methodology

2.3.1 Field Collection

Field activities included three distinct sessions of data collection from late fall 2004 to early spring 2005. The first collection period occurred November 17 and 18, 2004 at

Glen Haffey and Crawford Lake respectively, and consisted of a conventional site survey. This undertaking included the collection of terrain information using a Trimble 5700 survey-grade (2 cm vertical and 1 cm horizontal accuracy) GPS receiver. We collected 1094 data points over the sub-hectare Glen Haffey site and 10611 points over the Crawford Lake site. Simultaneous to this terrain survey, standard Campbell Scientific Inc. CS615 and CS616 TDR were placed at four points in each site, inserted to 30cm. Each instrument was inserted vertically to provide an averaged soil moisture content at each point. The TDR probes have a support of 300 cm^3, and the extents of the nano-catchments are 1786 m^2 and 6233 m^2, while the spacing of the TDR probes varied from 6 m to 26 m and 4 m to 30 m for Glen Haffey and Crawford Lake, respectively (Blöschl and Sivapalan, 1995). Manual single-ring infiltrometers fed by Mariotte bottles were placed throughout the sites and monitored regularly to provide reference data for future infiltration modelling.

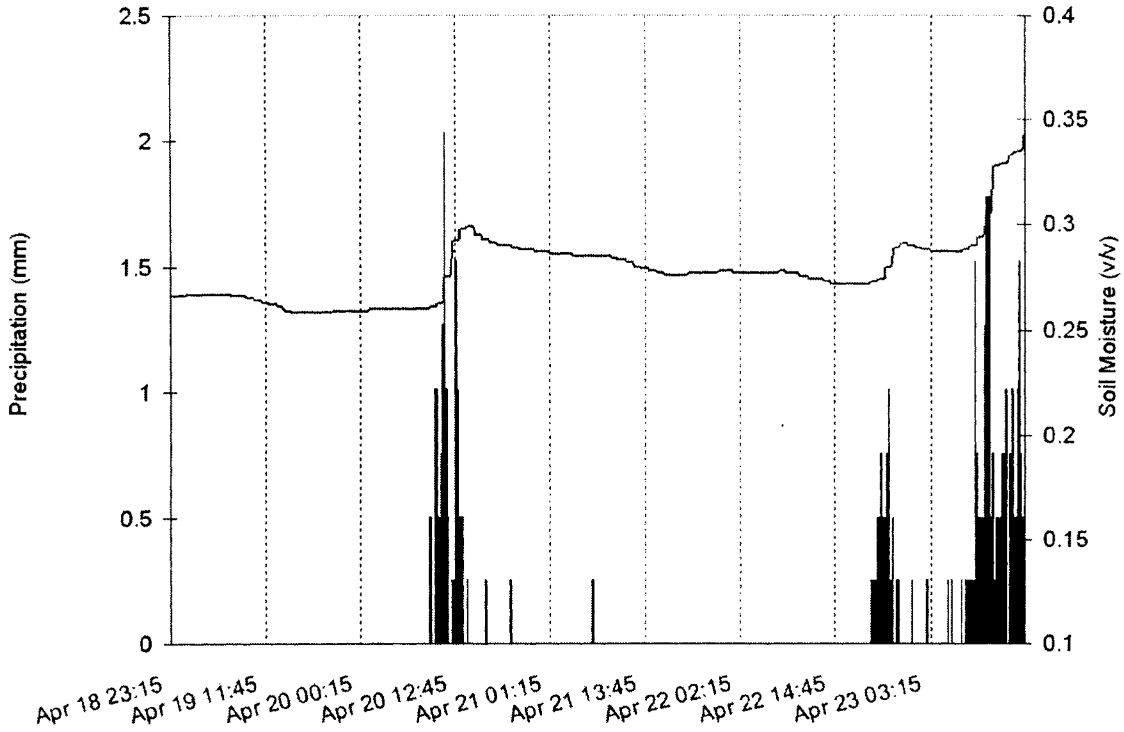

Figure 2.3 Precipitation (bars) and soil moisture content (line) as measured at the Glen Haffey meteorological station during the period April 18 to April 23, 2005

The second session took place at the Glen Haffey site from April 19 to April 23, 2005. Rain for the ten days prior to the initiation of field collection was minimal, with no precipitation falling the three days prior to April 19. Precipitation events occurred sporadically throughout April 20, 21 and 23, while April 19 and 22 provided fairly stable hydrologic regimes (Fig. 2.3). This session used standard techniques which focused on the use of statically located TDR sensors, and also incorporated the use of mobile TDR sensors. Through the use of the ProbeFusion data acquisition system (Graniero and Miller, 2003), designed to integrate environmental sensors and GPS, two mobile TDR were used in conjunction with a CS107 temperature sensor and a SDEC Inc. SKM 850t mobile vacuumeter to monitor several tensiometers installed across the site. Over this five day field campaign, soil moisture measurements were made at points that averaged the top 15 cm and 30 cm of the profile. Between April 19 and 22, 962 and 904 measurements were made at different points for the 15 cm and 30 cm depths, respectively. However here we are only considering data from April 19 and 22 as a result of the fact that precipitation events during other days of field collection lead to samples that were too temporally variable. For each sample point two TDR probes were manually inserted to their respective depths. The manual insertion presented the possibility of bending the waveguides, which was minimized through the use of an insertion tool and careful deployment. ProbeFusion allowed for abnormal data points to be detected during collection, and the probes were re-inserted when issues arose. The sampling strategy for soil moisture monitoring was dependent on the observations of the fieldworker. Thus a quasi-stratified random sampling strategy was undertaken in order to insure complete coverage of the site. A slight clustering in the final dataset occurred for two reasons. First, slightly more samples were taken near the six static tensiometers, soil moisture measurements are at these sites required in order to cross-reference against matric potential for infiltration modelling. Second, inserting the TDR probes was more problematic in some areas, and therefore the areas were slightly under-sampled.

The final session occurred at Crawford Lake June 14-15, 2005. We followed the same general methodology as we used during the prior acquisition period. During this period 173 samples were collected at each depth. This was far fewer than would we would have anticipated collecting, and was specifically a result of equipment issues. This small sample is less than would be ideal to order to perform full geostatistical analysis (Western et al., 1998b), but was deemed sufficient for the majority of analysis performed in this study. With no rainfall the hydrologic regime proved to be fairly stable during this collection event.

2.3.2 Analysis

The analysis of the data includes both analyses of the pattern soil moisture and generation of this pattern in terms of geostatistics. Several parameters expected to influence soil moisture were studied in terms of their relation to the resultant moisture pattern. These will be discussed in greater detail below, but include differences in vegetation / leaf litter and topographically derived variables including slope and wetness index.

2.3.2.1 Soil Moisture Pattern

Variography was used to analyze the pattern of soil moisture and the spatial auto-correlation between points. The variogram is a tool which describes the degree of similarity between two points with respect to their separation (Western et. al., 1998a). Several variogram models were initially studied, including the spherical, Gaussian and exponential model. After examining the resultant semi-variance, standard error and root-mean square error of the predicted models it was determined that the exponential model served (Eq. 1 and Fig. 2.4) fit best,

$$\gamma(h) = 1 - \exp\left(\frac{-h}{a}\right) \tag{2.1}$$

where $\gamma(h)$ semivariance with respect to the separation distance h between points (m) and a is the range (m) beyond which the semivariance between points does not effectively change. The sill of the semivariogram is the amount of semivariance where spatial auto-correlation becomes almost non-existent. The use of the exponential model has been used in several studies with a non-zero nugget (the semivariance at an infinitely small separation distance) used in some instances (Western et al., 2004), but not in others (Western and Blöschl, 1999), both with varying degrees of success. In this study a non-zero nugget was used, as it allows for some assessment of the variability that occurs within measurements that are very close together and that results from instrument error and finer-scale patterns. The nugget is difficult to explain with any certainty. These parameters help describe the degree of organization in the pattern of soil moisture. As mentioned the soil moisture pattern is assumed to be more randomly distributed during drier periods and therefore lower ranges would be expected. For each variogram model a number of bin sizes and number of lags were studied using a simple sensitivity analysis.

Where the sill and range varied minimally and RMS was low the bin size and the number of lags were selected.

The temporal variation across and between the two soil depths of study was also explored using variography. We explored both the change in the range through time and between depths, while also taking into account for the variability of the sill. Therefore the focus in terms of the variogram analysis was primarily the range and sill.

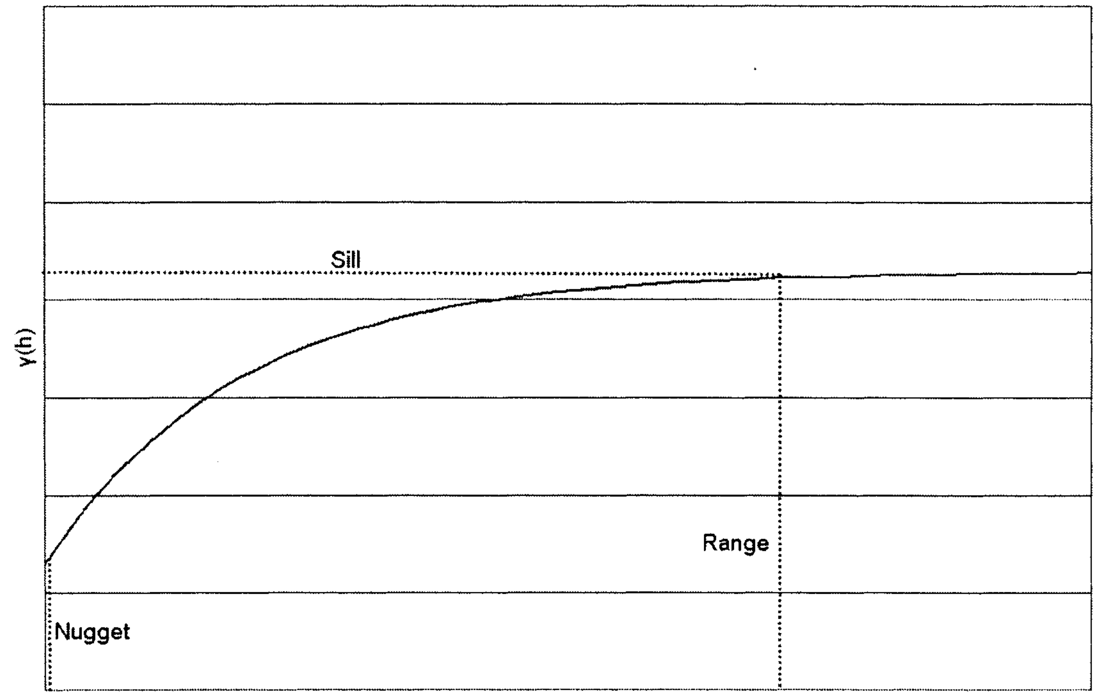

Figure 2.4 Theoretical exponential variogram curve

A digital elevation model (DEM) was created for each site using all 1094 points at Glen Haffey and 4164 points at Crawford Lake for each model. The DEM was generated through ordinary kriging using the Gaussian model to fit the variogram curve. The Gaussian algorithm was selected as it provided the most accurate model of the local topography when compared to both the spherical and exponential models. The resolution of these elevation models, were 0.25 m Glen Haffey and 0.34 m Crawford Lake. The DEM was used to calculate the local slope and wetness index (Eq. 2, Beven and Kirkby, 1979) for each cell in the bowl.

$$WI = \ln\left(\frac{a}{\tan\beta}\right) \qquad (2.2)$$

where a is the local upslope area draining through a given point per unit contour length and $\tan\beta$ is the local slope. The wetness index was smoothed by averaging over a 3*3 window in an attempt to minimize any issues arising from the fact that the scale of this study is an order of magnitude smaller than most studies using the wetness index (e.g. Brasington and Richards, 1998; Wolock and McCabe, 2000). As the resolution of the DEM has an effect on the topographic indices generated (Brasington and Richards, 1998; Wolock and McCabe, 2000) we coarsened the original DEM. However, no greater information could be extracted with specific relation to the wetness index. This coarsening results in the slope calculated to decrease and the contributing area to increase (Brasington and Richards, 1998), thus minimizing some of the effect of changing the resolution on the resulting wetness index derived.

The presence or absence of leaf litter and differences in the vegetation cover type were mapped (Fig. 2.2). These polygons were used to spatially mask the soil moisture data into sub-sets for comparison. Standard regression analysis was performed for this on each dataset, and on each sub-set defined by vegetation cover. The strength of the relationship between soil moisture and the topographic measures was evaluated. The correlation between soil moisture averaged over 15 cm and 30 cm was also evaluated by assessing box-plots derived from the various datasets.

The soil moisture values were transformed to z-scores in order to better compare the relative degree of wetness over time (i.e. temporal persistence). The distributions sufficiently fit a normal distribution (Kolmogorov-Smirnov test: $p>0.15$ for both April 19 and 22 at 15 cm measures; $p>0.149$ for April 22 at 30 cm; $p<0.1$ for April 19 at 30 cm) to support the transformation. These transformed values were used to assess relative wetness at a point over time. The relative scale allows for the comparison of these surfaces through time to monitor in the short term whether areas within the bowl maintain any consistency in their relative soil moisture over a rainfall event (in the case of Glen Haffey). Areas that consistently represent the bowl's average and extreme soil moisture contents may also be distinguished (Vachaud et al., 1985; Gomez-Plaza et al., 2000; Pachepsky et al., 2005). An interpolated surface of soil moisture distribution was created using kriging and the same model type as described for raw soil moisture values including the same bin sizes and number of lags.

2.4 Results

The initial results from the autumn of 2004 provided the basis for further investigation of soil moisture variability within both Glen Haffey and Crawford Lake as both data sets provided sufficient evidence that soil moisture varied spatially throughout the respective sites. Within the Glen Haffey study area a 23% difference in volumetric soil moisture content was observed across 28 m. This variability was also exhibited at Crawford Lake with soil moisture content ranging over 20% across 30 m. While soil moisture varied considerably through space, temporally the soil moisture was nearly static over the day. Therefore measurements collected over a single day could be assumed to be representative of the same sample in the absence of precipitation.

2.4.1 Spatial Distribution

The variograms of soil moisture revealed a number of spatial relationships (Fig. 2.5). Soil moisture at this study scale has a relatively short range (i.e. distance over which there is significant auto-correlation) as a result of the numerous confounding factors which influence its pattern. The correlation length for soil moisture within the Glen Haffey nano-catchment was 14.82 m at 15 cm and 14.66 m at 30 cm for April 19. The same measures on April 22 were 19.36 m and 19.62 m, respectively. This increase in correlation length is tied to the relative increase in soil moisture content within the bowl that resulted from several prolonged precipitation events during the period from April 20 to April 22 (Fig. 2.3). The similarity of range lengths at the respective depths indicates that there is good vertical continuity in the top 30 cm and that similar processes are acting on the soil moisture pattern throughout that layer. The correlation lengths from the soil moisture variogram for Crawford Lake are much greater at 44.66 m for both 15 cm and 30 cm depths on June 14, 2005. The data gathered for the June 15 sampling date were too sparse to generate a satisfactory variogram. The difference in correlation length between Crawford Lake and Glen Haffey can be interpreted in a number of ways, including that the overall pattern of soil moisture has greater spatial connectivity at Crawford Lake. However, the relative difference in the size of the nano-catchments and the resulting sampling density must be considered. The sample spacing is smaller at Glen Haffey, therefore there is a better chance of detecting short-scale variability. As one moves from smaller to larger study sites, sampling generally becomes sparser and thus variability can be missed, as is observed in an increase in the variogram range at Crawford Lake. This difference highlights the scaling issues that propagate the literature on the analysis of soil moisture patterns. Other scaling controls include macropores at relatively fine scales and progressively change as the resolution of the study becomes coarser to include large topographic feature, vegetation, etc. (Wilson et al., 2005).

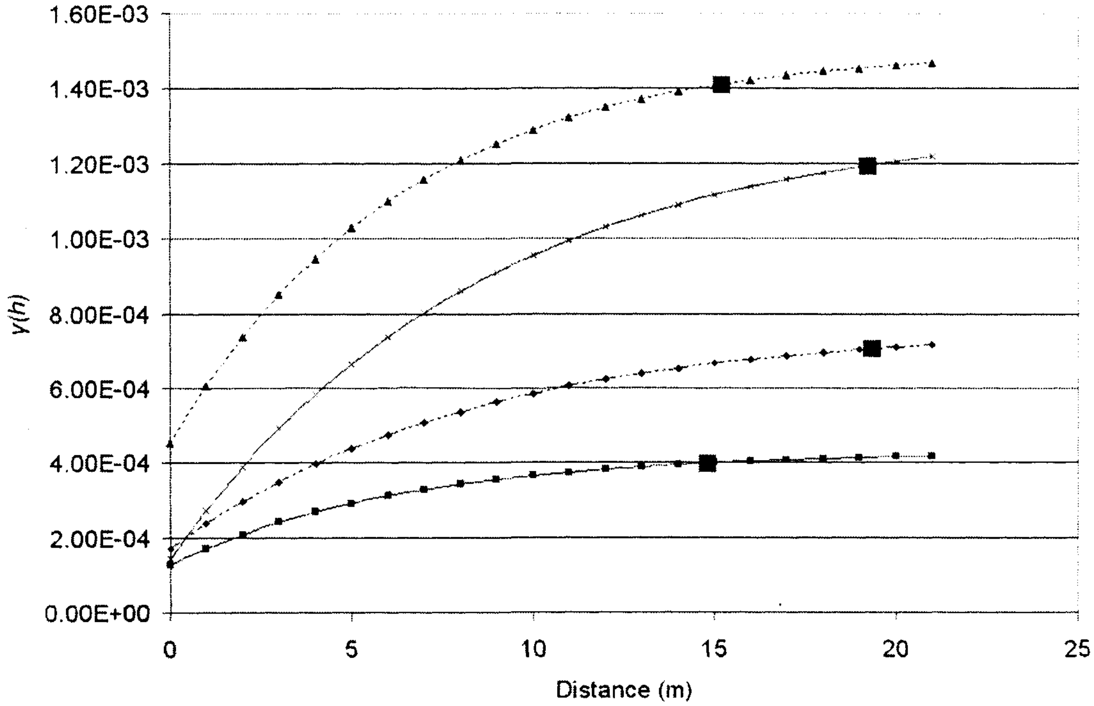

Figure 2.5 Variogram of soil moisture within the Glen Haffey nano-catchment. Solid line with squares – April 19, 15 cm; Broken line with triangles – April 19, 30 cm; Broken line with diamonds – April 22, 15 cm; solid line with cross-hatches – April 22, 30 cm; Large squares – variogram range

In assessing the use of a z-score transformation we are limited slightly because of our data. However some observations can be made. The spatial z-score distribution at Glen Haffey indicates that those areas which are at the highest elevations are generally below the mean soil moisture, whereas those areas that are proximal to the deciduous forest to the west or are in the lower portion of the bowl are higher than the mean (Fig. 2.6). Within the area closer to the deciduous forest, leaf litter serves as a control which homogenizes the variability of soil moisture in the area. This also provides evidence that elevation alone does not provide a satisfactory control on the resultant pattern of soil moisture. Even within drier portions of the bowl elevation alone cannot be deemed to be the sole control as those sites which are on either west- or south- facing slopes are prone to be dry versus those areas which face east or north. However, aspect does not appear to be a factor everywhere. The pattern of relative soil moisture is broadly stable between study days, though the average "zero line" does fluctuate slightly. With a more temporally extensive dataset the use of a z-score transformation shows promise in elucidating the pattern of relative wetness within a nano-catchment.

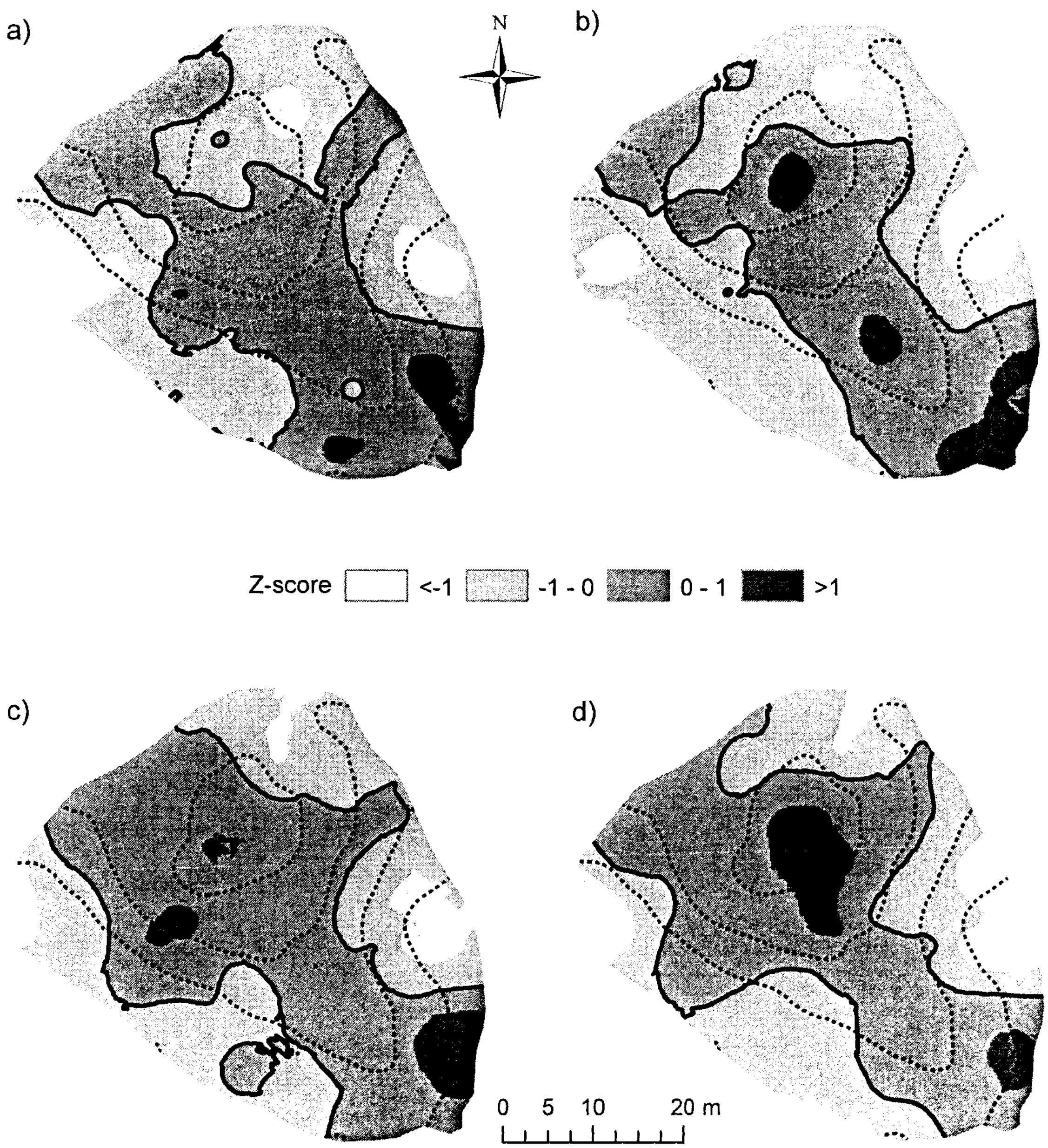

Figure 2.6 The z-transformed soil moisture distribution at Glen Haffey. A) April 19, 2005 at 15 cm b) April 19, 2005 at 30 cm c) April 22, 2005 at 15 cm d) April 22, 2005 at 30 cm. The solid black line is representative of the average 'zero line' while broken lines are elevation countours are at 1 m intervals.

2.4.2 Topographic Controls

Topographic measures showed little or no visible control on the pattern of soil moisture outside the lowest elevations within the bowls. Both surficial and sub-surface flow paths converged at the bottoms of the nano-catchments generating a higher soil moisture content. The bowls do not seem to be large enough to give much range in the wetness index values following the conventional calculation method. Given the geometry of these bowls, the index only distinguished points where convergence was at a maximum or at the highest relative elevations (Fig. 2.7). In the areas where topographic convergence was not as strongly defined, topography on its own was not a sufficient control to describe the resulting soil moisture pattern.

As expected, figures 2.8 and 2.9 show that the relationship between terrain, vegetation and soil moisture is complex. While no strong statistical elevation trends are present, separating the vegetated / non-vegetated areas at Crawford Lake and the leaf litter / no litter areas at Glen Haffey produced distinctly different trends in each group. At Crawford Lake a relatively weak relationship is discernible; with increasing elevation there is a decrease in soil moisture (Fig 2.8). This pattern is amplified where a vegetative cover exists, as the coefficient of determination (r^2) for all four data sets at a minimum doubles in areas covered by vegetation. r^2 for June 14 at 15 cm is 0.183 for bare ground sites, while for vegetated areas the r^2 is 0.705. This pattern as mentioned is seen in all data sets, but may be most intriguing for the dataset of June 15 at 30 cm (Fig. 2.8 d), where for bare ground almost no relationship between topography and soil moisture (r^2 = 0.071), while on vegetated areas r^2 is 0.483. While not statistically significant, a clear pattern is present in the vegetated dataset that is absent in the open areas of the nano-

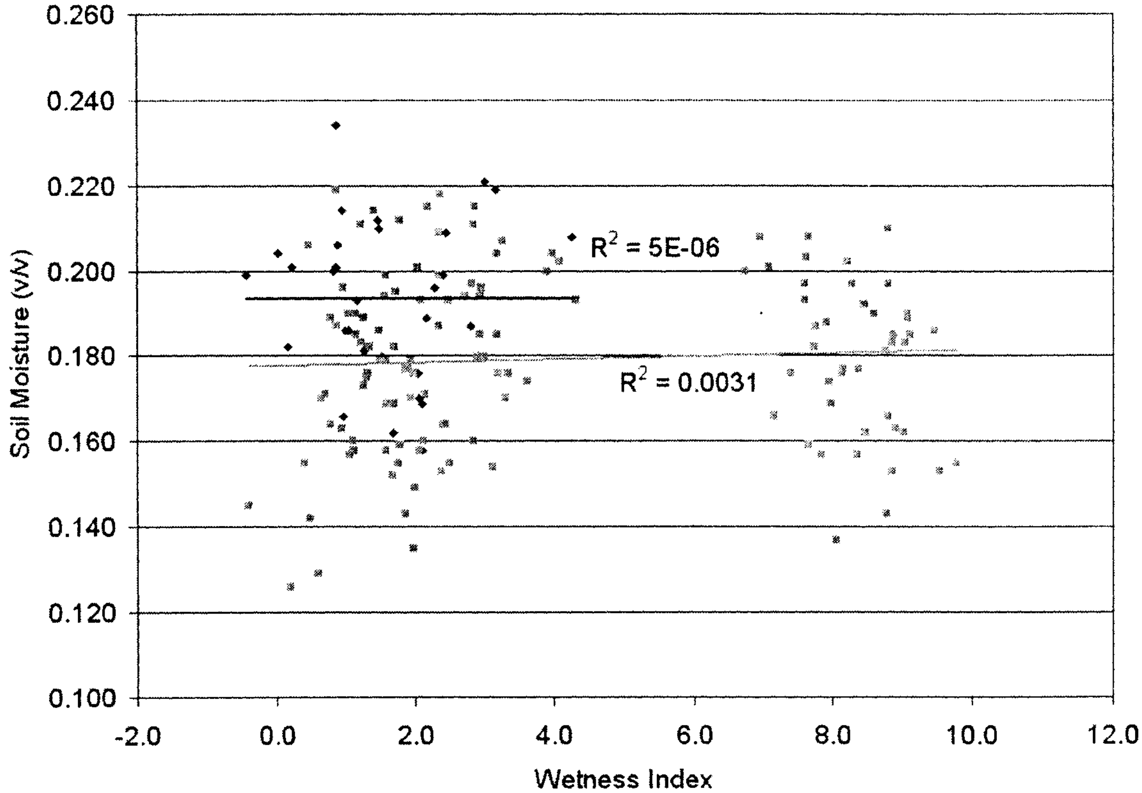

Figure 2.7 The relationship between the computed topographic wetness index and soil moisture at 15 cm on April 19, 2005 within the Glen Haffey study area. Grey squares – no leaf litter; Black diamaonds – leaf litter present

catchment. Within mid-elevation sites of the bowls a clustering of soil moisture occurs, but generally soil moisture is slightly less in vegetated areas.

At Glen Haffey, where leaf litter is present soil moisture remains more consistent and elevation as a control is minimized as r^2 is less than 0.1 for all four datasets (Fig. 2.9). When leaf litter is absent there is an observable relationship between soil moisture and elevation with the exception of the April 19, 15 cm dataset ($r^2 = 0.089$). This data set is also interesting because not only does the leaf litter minimize any potential effect of elevation on soil moisture, but indeed reverses what would generally be assumed as soil moisture increases with elevation ($y = 0.0122x - 4.9$) (Fig 2.9 a). This same pattern again presents itself in the data at 15 cm on April 22 (Fig 2.9 c), where leaf litter again serves as a control against micrometeorological forcing. Leaf litter is spatially confined as it is not present at lower relative elevations. This is result of the proximity of the western margin of the bowl to the deciduous forest stand. The effect of leaf litter at 30 cm is minimal (Fig 2.9 b and d) as surficial controls are suppressed with depth.

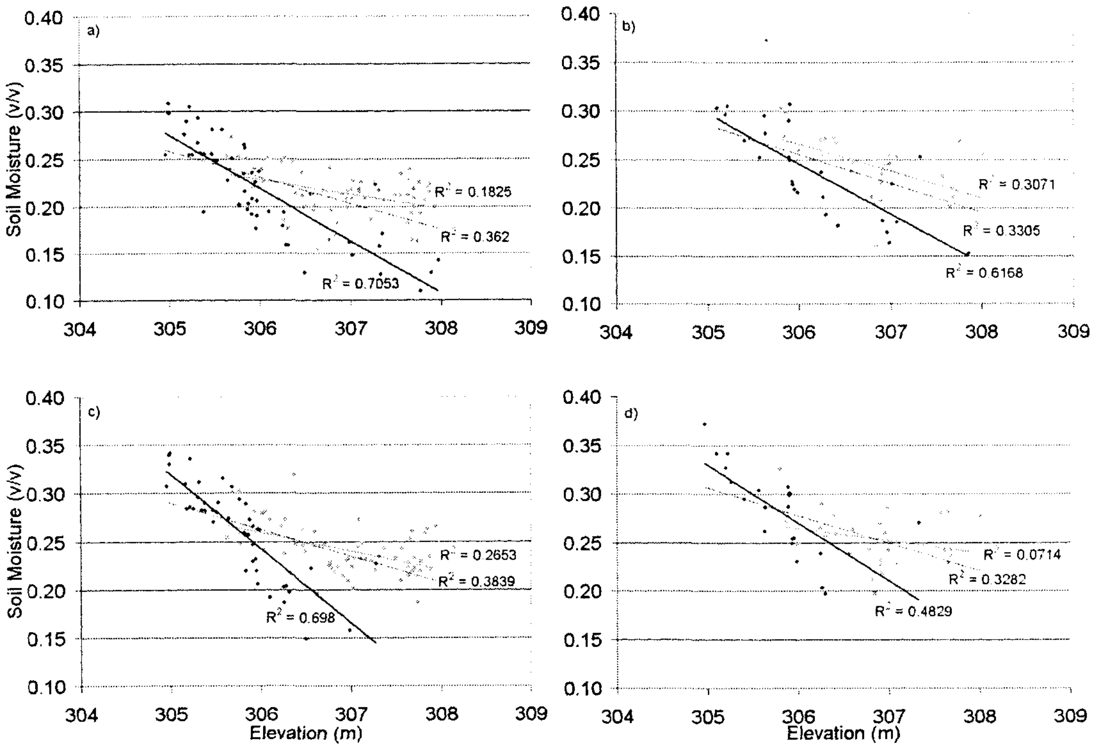

Figure 2.8 Relationship between soil moisture and elevation at Crawford Lake field site a) June 14, 2005 at 15 cm b) June 14, 2005 at 30 cm c) June 15, 2005 at 15 cm d) June 15, 2005 at 30 cm. Grey squares – bare ground; Black diamonds – vegetated; solid broken line – trend line for measures under active vegetation; solid gray line – trend line fallow area; broken line – trend line for entire dataset

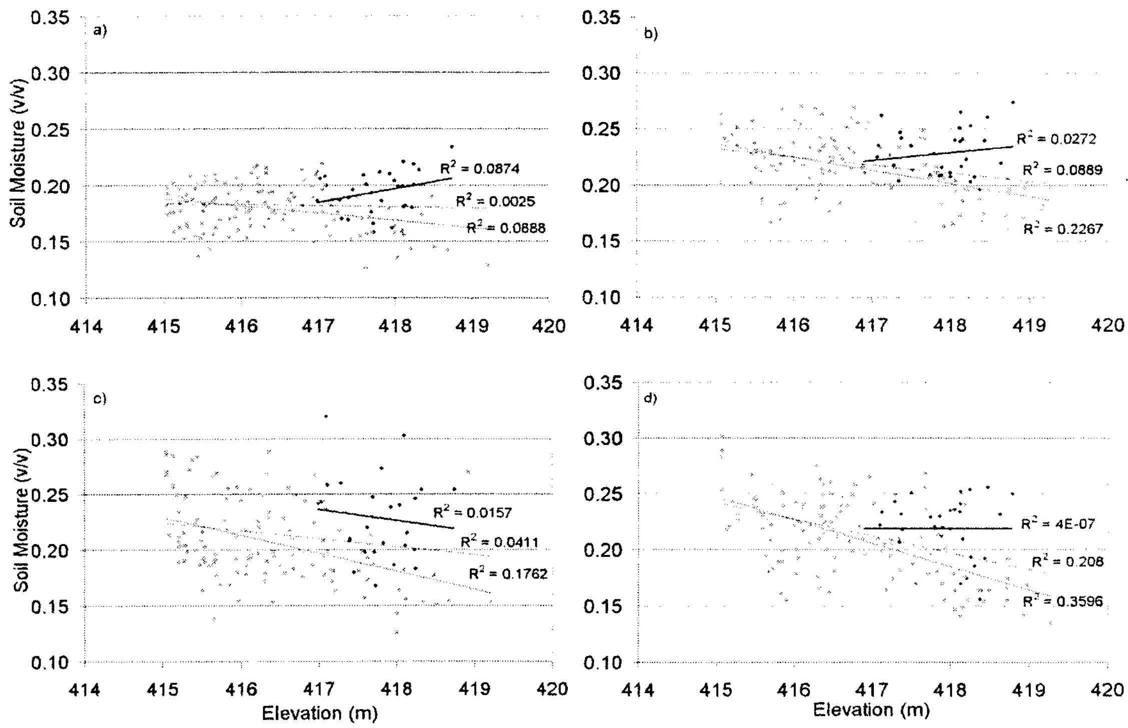

Figure 2.9 Relationship between soil moisture and elevation at Glen Haffey field site a) April 19, 2005 at 15 cm b) April 19, 2005 at 30 cm c) April 22, 2005 at 15 cm d) April 22, 2005 at 30 cm. Grey squares – bare ground; Black diamonds – vegetated; solid black line – trend line for under leaf litter; solid gray line – trend line for grass only measures; broken line – trend line for entire dataset

Additionally leaf litter serves as a barrier to evapotranspiration at Glen Haffey, thereby minimizing the variability of soil moisture in the upper 30 cm of the soil profile when compared to those areas which do not have this O_i horizon (Fig. 2.10). Those areas under this cover are also moister than are those exposed areas prior to and after precipitation events, which would be anticipated. Comparatively, soil moisture at 30 cm during both 'dry' and 'wet' conditions varies considerably more than soil moisture at 15 cm. This is fairly counter-intuitive as both percolation and evaporation would be anticipated to occur at faster rates in the very near surface and would normally vary more spatially as a result of aspect most especially.

The effect of leaf litter on near surface soil moisture is a great deal less than that of vegetation in comparison to bare ground. However, this should be expected as even where leaf litter is present vegetation transpires and serves as a homogenizing agent on the overall Glen Haffey nano-catchment. At Crawford Lake the dichotomy between vegetated and bare ground is remarkably apparent and again comes down to the effect of transpiration in water-stressed areas of the bowl and as a shade against evaporation in areas that have sufficient moisture. Thus overall vegetation serves as an amplifier of topography within the Crawford Lake site, while leaf litter minimizes the effect of terrain.

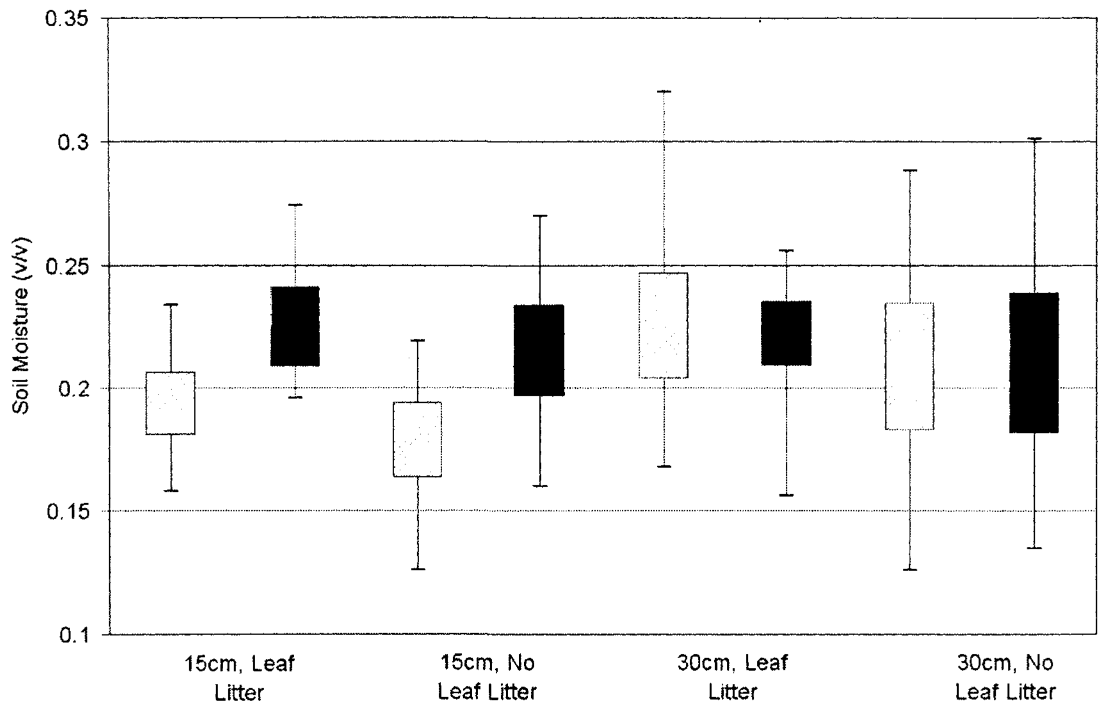

Figure 2.10 Soil moisture distributions for varying land covers at Glen Haffey on April 19 (Grey) and April 22, 2005 (Black)

2.5 Discussion and Conclusion

The pattern of soil moisture over a small area is controlled by a variety of factors working at numerous simultaneous scales. The overall pattern of soil moisture exhibits a significant degree of variability, with correlation length (variogram range) between proximal points being at a minimum during dryer periods. The variogram range at Glen Haffey varied from 14.6 m to 19.6 m, depending on the depth, which is low compared to 44 m at Crawford Lake. Lookingbill and Urban (2004) for example found a range of 50 m, while Grayson and Western (1998) document ranges from 10 m to 1000 m depending on the size of the study area. The ranges at our sites fit fairly well into this context. An increase in correlation following precipitation input is also consistent with other studies (Schume et al., 2003; Western et al., 2004; Wilson et al., 2005). This is a result of an increase in the spatial connectivity throughout the bowl. The connectivity is extensive because no distinct gullies are found in either site, which would lead to a spatial discontinuity and affect the resultant correlation lengths.

In this study we see evidence of measurement scale affecting the soil moisture pattern as illustrated by the short correlation lengths of the variograms. However in quantifying the resulting variability it is difficult to account for the controlling agents. The study of several topographic attributes and their subsequent relation to the pattern of soil moisture provided little insight into point measures of soil moisture within the bowls. However, this is juxtaposed in areas of highest topographic convergence (i.e. lowest portions of the bowls) where soil moisture content is greatest within both sites. It has been assumed that topographic gradient provides a significant control on the soil moisture pattern within an area, especially during wet periods (Green and Erskine, 2004).

However, no clear relationship was established within the intermediate and higher elevations using topographic indices alone at our sites, even during 'wet periods'. The wetness index, often used for determining zones prone to extreme soil moisture conditions (Woods and Sivapalan, 1997; Western et al., 1999), was not a sufficient predictor here. This arises from the very small site size compared to those catchment studies where the wetness index is a suitable predictor. At the nano-catchment scale the operative processes governing lateral redistribution are not adequately represented in the wetness index formulation, and here is not a suitable predictor for soil moisture.

While topographic measures alone did not provide sufficient insight into the resulting soil moisture pattern, the use of both terrain and ground cover did provide some explanation of the pattern of soil moisture. Distinct differences in the relationship between elevation within the nano-catchment and soil moisture under a given cover type was displayed for both the Glen Haffey and Crawford Lake sites. This divergence was particularly strong at Crawford Lake where the vegetation / no vegetation difference was more extreme. The enhanced gradient can be assumed to be a result of both topographic gradient and plant transpiration. The vegetation effect is superseded at lower elevations where topographic convergence plays a much greater role. At higher elevations at Crawford Lake there is a bedrock layer which precludes the insertion of the TDR instruments to 30 cm in some areas, however the heightened inverse relationship is still observable at the upper limit where both 30 cm and 15 cm measurements could be made.

The effect of leaf litter on the homogeneous grass surface at Glen Haffey is less prominent than the vegetation differences at Crawford Lake, though a sufficient difference arises in areas where leaf litter is present that it should be accounted for. Leaf litter, unlike living vegetation, does not actively remove water from the matrix but instead serves as a barrier against evapotranspiration from the soil and vegetation below this temporary cover. Actually accounting for leaf litter in monitoring soil moisture is difficult because of its sporadic nature. Presumably the final resting place for litter will involve the interaction of local wind eddies with the bowl morphology.

In addition to assessing the variables associated with the spatial pattern of soil moisture, a significant amount of potential exists in the use of z-score transform for assessing the relative soil moisture pattern at any location through time. Within this study we were limited to a small temporal sample though with a more rigorous field campaign the z-score transform presents a number of possibilities in studying the temporal stability of soil moisture within a nano-catchment. This includes the assessment of a location within a catchment which has an average soil moisture content through time (Grayson and Western, 1998) and with a greater amount of data this measure could be very useful in monitoring the stability of the pattern of soil moisture through time (i.e. seasonality of relative soil moisture) (Gomez-Plaza et al., 2000). So, future work will attempt to determine whether the use of a simple z-score transform would aid in clarifying any potential temporal stability within these sites.

It is very difficult to separate topography and ground cover in assessing the resulting pattern of soil moisture (Grayson and Western, 2001; Ridolfi et al., 2003). As greater detail becomes increasingly available through LIDAR and satellite imagery on both terrain and land cover it is not clear that one is more important than the other for predicting the soil moisture pattern at this current scale of study. While it would be ideal if one variable could be used as an indicator of the soil moisture pattern, this is clearly not possible at these sites. At Crawford Lake active vegetation heightens the basic relationship between soil moisture and topography. At Glen Haffey the presence of leaf litter creates a more uniform soil moisture, overriding the weak topographic relationship. This does suggest that the flat locations of most regional monitoring stations do not likely create measurement bias with respect to the highly variable microtopography of the ORM. Separating vegetation cover made it easier to identify localized soil moisture trends. This indicates that high resolution land cover (IKONOS, Quickbird) data would be beneficial for identifying areas which maybe moister than predicted at regional stations. However, a more precise estimate of the higher soil moisture would be difficult without the LIDAR-style terrain data. The full role of other properties including soil texture, micrometeorological attributes and bedrock features have not been evaluated yet, but would presumably strengthen these relationships. Therefore continued research is required to understand the pattern of soil moisture within and between field sites at the nano-catchment scale.

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CHAPTER III
On the scaling of infiltration: Spatial and temporal patterns of matric potential and infiltration to saturation in two sub-hectare 'nano-catchments'

The manuscript titled, "On the scaling of infiltration: Spatial and temporal patterns of matric potential and infiltration to saturation in two sub-hectare 'nano-catchments'" is co-authored by P. Andrew-McBride and P.A. Graniero and targeted for Journal of Hydrology. It has yet to be submitted as of January 13, 2006.

3.1 Introduction

The study of infiltration, the downward movement of water through the soil's surface, and the attributes associated with it is of increasing importance. As more 'natural' lands are anthropogenically altered through urbanization and farming practices, understanding both the rate of infiltration and the variability of this phenomenon is critical. The process of infiltration serves as the primary mechanism by which groundwater is recharged in a number of terrestrial environments (Becker and Frind, 2000). Infiltration also controls to a great extent the movement of pollutants in solution and suspension into local aquifers and via subsurface flow paths. Additionally, with increasing urbanization, infiltration is reduced significantly by surface seals created in the form of roadways and diverted by sewer networks.

The variability of hydrologic phenomena associated with infiltration as a result of a rainfall event over space and time has been well studied (Loague and Gander, 1990; Goodrich et al., 1995; Govindaraju et al., 2006), however the dynamics associated with this variability are not completely understood. This is especially important in studies attempting to model the hydrological dynamics within catchments and aquifers. Typical studies create hydrologic units represented by grid cells within the model (Jain et al., 2004). It is assumed that variability within these units occurs, though the degree of variability is generally unknown and is therefore ignored. Among the features that lead to variability within these cells are differences in vegetation, topography, soil texture, and amount of organic matter. It is the goal of this work to identify and explain some of the variability of soil moisture, matric suction, hydraulic conductivity, and the resulting infiltration up to saturation through both space and time during intermittent rainfall events and under simulated conditions. The quantity of water infiltrating prior to saturation is critical in estimating the amount and timing of the onset of potential ponding and/or overland flow.

3.1.2 Background

Measuring the hydraulic parameters associated with infiltration at either a point or on an areal basis can be costly and very time consuming. Therefore the use of pedotransfer functions (Bouma, 1989) that relate measured soil properties to the hydraulic characteristics of a unit under study has become a standard practice (Saxton et al., 1986; Castillo et al., 2003). These pedotransfer functions are used in determining properties associated with the process of infiltration directly and also in determining the attributes of the soil water characteristic curve. The soil properties which are normally used to derive the hydraulic variables include soil texture, porosity (ϕ), organic matter (OM), organic carbon, field capacity (θ_{fc}), saturated soil moisture content (θ_s) and residual soil moisture content (θ_r) (Rawls et al., 1983b; Wosten et al., 2001; Fredlund et al., 2002; Nemes et al., 2003; Teepe et al., 2003; Tomasella et al., 2003; Rajkai et al., 2004). Databases of these properties are becoming increasingly prominent as the use of pedotransfer functions increases and groups of soil scientists integrate their respective datasets (e.g. UNSODA and HYPRES) (Nemes et al., 2003). Within most hydrologic modelling studies soil attributes are derived from medium-scale soil maps (Lin et al., 2005). Therefore fine-scale heterogeneity is lost, both in terms of soil properties and the resulting hydraulic characteristics. The process of infiltration, while generally controlled by the textural properties of the soil, is greatly influenced by a number of other factors, including topography, vegetation and macropores (Flint et al., 2002; Santos et al., 2003; Weiler, 2005).

Topography affects infiltration at a site as both microtopography and larger topographic features act as partitioning agents as rainall reaches the surface. This effect includes the spatial redistribution of soil moisture and erosional effects on the soil surface (Santos et al., 2003). While a limited but continually growing amount of work has been done on the role of topography on hydraulic properties the "published results demonstrate some strong correlations" (Pachepsky et al., 2001). Topographic features such as very steep slopes serve to seal the surface regionally, as higher slopes and therefore gravity leads to an increase in overland flow and effectively eliminates infiltration at these sites. Microtopography, for example hummocky terrain, can have the same type of influence on localized infiltration (Hansen et al., 1999; Hansen, 2000). Hummocks and other fine-scale topographic features have an effect on the detention storage capacity, which alters the temporal dynamics of infiltration by forming small ponds at times when runoff would be expected. It is therefore important to monitor changes in matric suction and soil moisture content as parameters with changes in topography which alter the variability of infiltration rates. Topographic indices have also been used in the study of water retention (Grayson and Western, 1998; Pachepsky et al., 2001; van Wessemal et al., 2003). These indices include slope, curvature, aspect, and wetness as evaluated by the wetness index (Beven and Kirkby, 1979), which integrates slope and upslope catchment area. The relationship of soil moisture and topographic relief, in particular, has been the focus of much research, with varying degrees of success (Grayson and Western 2001, 1998; Andrew-McBride and Graniero, in review).

Macropores, which are caused by root activity, desiccation, macrofauna and freeze-thaw action (Buttle and House, 1997), lead to a great deal of spatial variability in the process of infiltration. These pores are generally difficult to account for, given their semi-random distribution. Within these channels in the soil matrix capillary potential is negligible and the downward flow is a result of gravity (Bronstert and Plate, 1997). Macropore flow is seen to occur only when the soil surface is near saturation. In infiltrometer experiments the effect is observable in the increased flow rates under saturated conditions that results from the required ponded conditions of the test. These macropores can lead to significant spatial variability in the rate of infiltration by altering the hydraulic conductivity, K [L/T], and the soil moisture content, θ, at depth. The stability of the wetting front is also influenced as a result of macropores, as preferential flow paths conduct water to greater depths in advance of the wetting front in the matrix (Weiler and Naef, 2003). One can assume that the presence of vegetation can be used as a proxy for the presence of these channels as a result of root activity, but quantifying the effect is more difficult. Characterizing the effect of macropores is complicated without a great deal of effort such as the use of dye tracer experiments or laboratory capillary tubes measurements (Sullivan et al., 1996; Weiler and Naef, 2003; Weiler, 2005).

3.2 Study Site Characteristics

The study areas are both situated in south-central Ontario, Canada (Fig 3.1). The sites are approximately 50 km apart, with the Crawford Lake study site situated south of Glen Haffey.

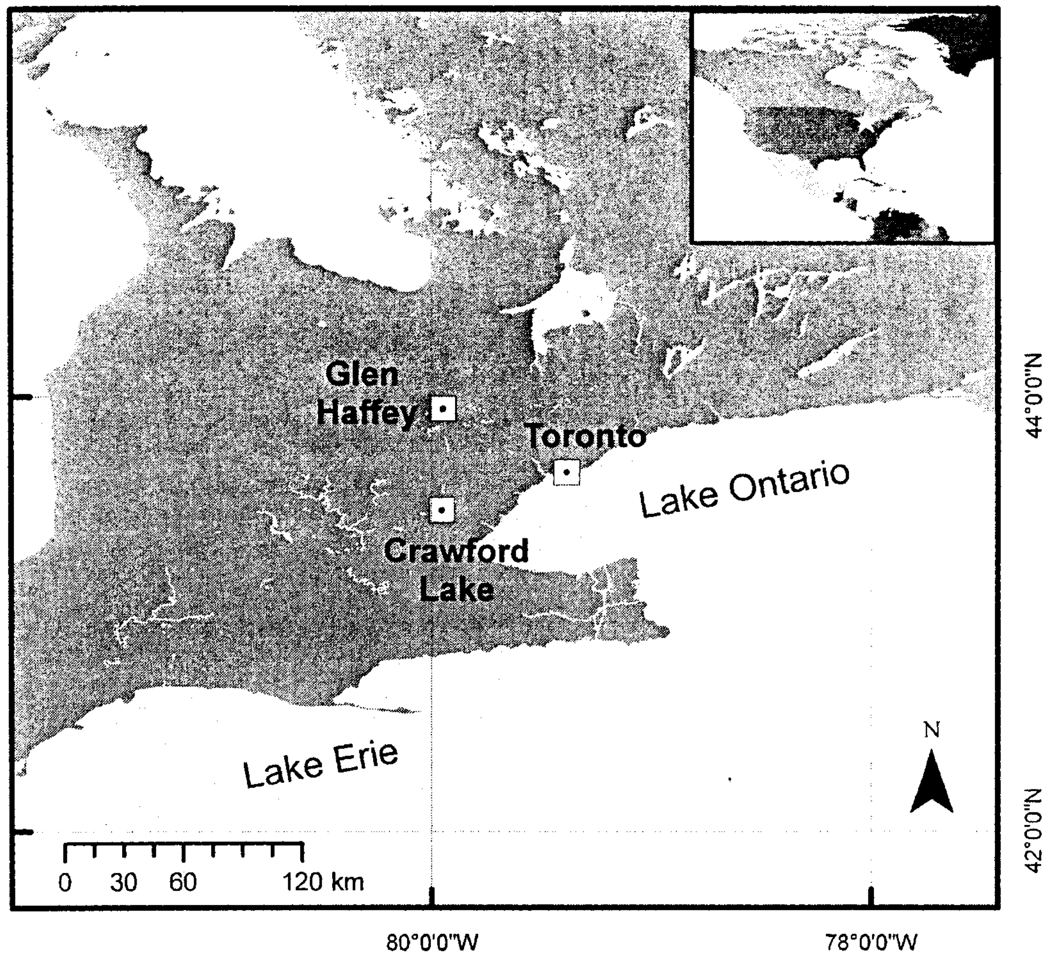


Figure 3.1 Locations of field sites in southern Ontario, Canada

3.2.1 Glen Haffey

The Glen Haffey field site is composed of a poorly sorted till that resulted from the Laurentide Ice Sheet. This area makes up a small portion of the Oak Ridges Moraine (ORM), which encompasses approximately 150 km wide glacial deposit (Barnett et al., 1998). The ORM serves as a major zone of aquifer recharge for a groundwater system that provides water for an approximate population of one million individuals. The ORM also serves as the headwaters for a number of tributaries, and the Glen Haffey area specifically serves as the upper boundary of the Humber River watershed. The grey-brown podzol is a sandy loam unit composed of an irregular steeply sloping area, with occasional pockets of gravel in the vicinity of Glen Haffey (Experimental Farm Service, 1953).

The study unit is approximately 1750 m^2 forming a 'nano-catchment', or sub-hectare bowl-like catchment, with no surficial outlet. The term was chosen to reflect that the sites are considerably smaller than those typically termed 'micro-catchments' by hydrologists'. The relief of the Glen Haffey site is approximately 4.5 m. The site's ground cover is composed of short grasses and clovers which are maintained throughout the growing season. The topography is undulating within the bowl and in the surrounding landscape, typical of this morainal landscape, with the slope varying considerably within the nano-catchment, as discussed below.

3.2.2 Crawford Lake

The Crawford Lake site is situated on the Waterloo aquifer and is underlain by the Niagara Escarpment, composed of a fissured limestone bedrock. This location, like Glen Haffey, is situated in an area of increasing urban growth, thus understanding the hydrology of the area is increasingly important. The site is itself is $6233m^2$ and has a relief of approximately 3.5 m. The vegetation of the area is divided into a near-care, fallow agricultural zone alongside a zone of successional native vegetation. The site is mapped as a grey-brown luvisol sandy loam (Canada Department of Agriculture, 1971), which is slightly stony and gently sloping (2-5%), however the purity (accuracy) of this map will be discussed later.

3.3 Methodology

3.3.1 Field Collection

Distinct data collection episodes occurred at each field site. One period involved the use of single ring infiltrometers to monitor infiltration across the field, while the other gathered intensive soil moisture and tensiometric data to evaluate the dynamism of these variables during sporadic precipitation inputs.

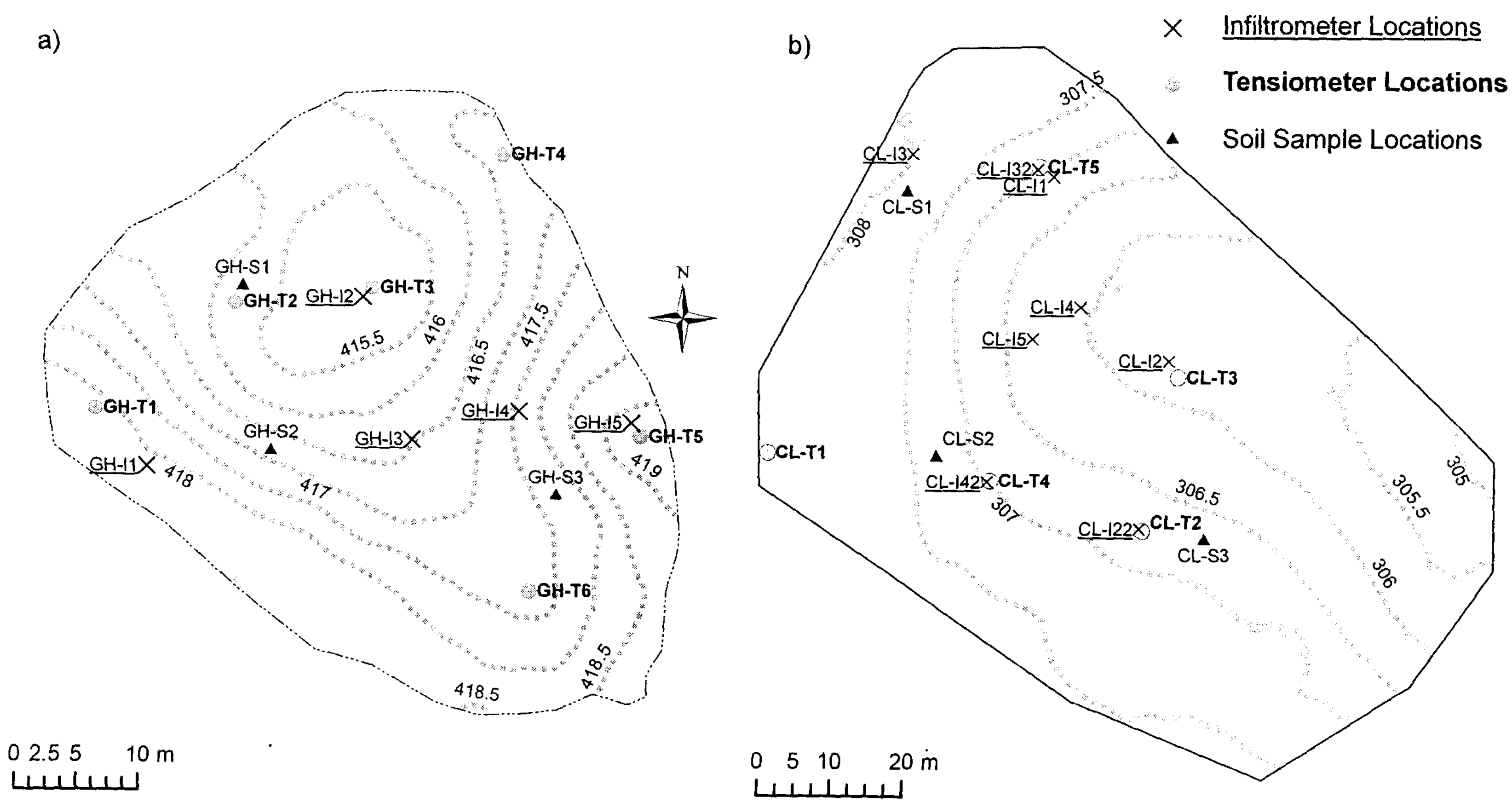


Figure 3.2 Distribution of sample locations at a) Glen Haffey; b) Crawford Lake

3.3.1.1 Infiltrometer Data

Single-ring infiltrometers were set up throughout both study areas to capture the infiltration rate at distinct points (Fig. 3.2). At Glen Haffey this analysis was undertaken in November 18, 2004 with five infiltrometers, while the same methodology was used on November 19, 2004 and June 15, 2005 at Crawford Lake, with five and three infiltrometers monitored, respectively. Each infiltrometer was fed by a Mariotte bottle in order provide a constant head of water at the surface (5 cm). The positions of the infiltrometers were situated in order to capture the greatest degree of variability with reference to both topographic position and local vegetation characteristics. While a horizontal flux of water below the base of the ring is assumed to exist (Buttle and House, 1997), this redistribution provides additional insight into the processes involved in the saturated flux of water in the near surface. This horizontal flux is especially important in areas that are on a sufficient slope. The infiltrometer data was used to estimate the saturated hydraulic conductivity K_s by using the Green-Ampt (1911) model:

$$\frac{1}{K_s} = \left(1 + \frac{\Delta\theta \cdot \psi_{wf}}{F}\right) / f \tag{3.1}$$

where f [L/T] is the rate of infiltration, F [L] is cumulative infiltration, $\Delta\theta$ is the soil moisture deficit (θ_s-θ_i), θ_s is the saturated soil moisture content and θ_i is the initial soil moisture content and where ψ_{wf} [L]is the suction at the wetting front as determined by pedotransfer function, to be discussed below.

3.3.1.2 Soil Moisture and Tensiometer Measurements

Concurrent to the monitoring of infiltrometers, soil moisture was assessed at four stationary points within the Glen Haffey nano-catchment November 18, 2004 and at Crawford Lake November 19, 2004, which provided short-term soil moisture during a period of no rainfall. These stationary points were monitored using standard Campbell Scientific Inc. CS615 and CS616 time domain reflectrometers (TDR), inserted vertically to 30cm.

During this two day period we also surveyed the nano-catchments. This undertaking included the collection of terrain information using a Trimble 5700 survey-grade (2 cm vertical and 1 cm horizontal accuracy) GPS receiver. At Glen Haffey we collected 1094 data points over the nano-catchment with a spatial density of 0.62 points per m^2. At Crawford Lake 4164 points were collected at a density of 0.67 points per m^2. These data were then used to produce digital elevation models (DEMs) with resolutions of 0.25 m and 0.34 m, respectively.

A second measurement session took place at the Glen Haffey site from April 19 to April 23, 2005. There was minimal rain during the ten days prior and no rain during the three days prior to April 19. Precipitation events occurred sporadically throughout April 20, 21 and 23 (Fig. 3.3). April 21 had a total of 0.254 mm of rain, which made no discernable change to the soil moisture or matric potential measurements. This session used standard, statically located TDR sensors as described above. Also, the ProbeFusion data acquisition system (Graniero and Miller, 2003) was connected to a Trimble 5700 GPS, two TDR (15 cm and 30 cm long) and a SDEC Inc. SKM 850t mobile vacuumeter to monitor local hydrologic characteristics. The mobile soil moisture samples were acquired over the respective depths (15 cm and 30 cm) by inserting the two TDR probes with different probe lengths vertically. An average of 186 points were sampled each day at Glen Haffey and 86 points at Crawford Lake, giving an average sampling density of 0.106 and 0.012 points per square meter, respectively. A quasi-stratified random sampling strategy, guided by the observations of the fieldworker, was undertaken in order to insure complete coverage of the site. General areas were revisited throughout data collection, but the temporal distribution of these repeat visits were not uniform and the precise sample locations were not necessarily revisited. Clustering occurred in TDR collection because of a bias in collecting measurements in proximity to the tensiometers.

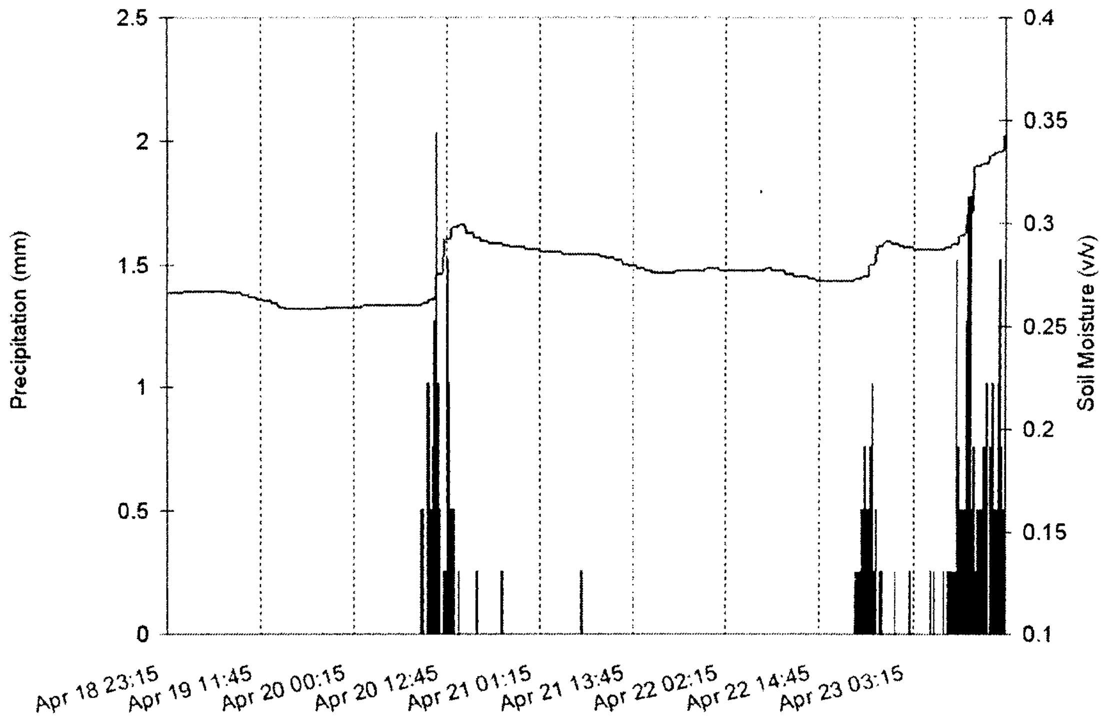


Figure 3.3 Precipitation and soil moisture content as measured at the Glen Haffey meteorological station during the period April 18 to April 23, 2005

The tensiometers were grouped into six sets installed at 15 cm and 30 cm, respectively, across the site. These clusters provide data to monitor the gradient of suction both vertically and horizontally. On each revisit, the mobile vacuumeter needle was inserted into the tensiometer membrane for a reading. However, due to equipment issues several of these tensiometers failed during this period, but a sufficient set of data were collected to provide information on the characteristics of matric suction. The tensiometers therefore provide information on suction under dry conditions and during the progression of the wetting front under infiltration.

Tru-check rain gauges were installed at each tensiometer cluster, and rainfall was measured manually to evaluate the fine-scale variation in rainfall amounts and rates. However, little variation in precipitation was discovered. In addition, rainfall rates were monitored at a weather station approximately 1.5 km from the Glen Haffey bowl. At this site the evolution of soil moisture during the precipitation period was evaluated at four depths (5, 10, 20 and 50 cm).

The Crawford Lake assessment of soil moisture was undertaken June 15 and 16, 2005. These two days were integrated into one dataset based on the similarity of the two data sets and the consistency of the hydrologic regime. This provided for a more robust soil moisture field. Tensiometric data were only gathered June 16, however this gives an initial indication of variability under 'dry' conditions. Again equipment issues arose, thus resulting in only three of five points having matric potential data at both depths.

3.3.2 Analysis

3.3.2.1 Soil Analysis

Soil samples of 5 cm in diameter and 5 cm in depth were gathered at three locations at the surface within each bowl, (Fig. 3.2). These samples were then used to ascertain a number of soil characteristics through a variety of standard laboratory procedures. This included textural analysis of surficial soils using standard sieving and pipette techniques (Black et al., 1965) to measure sand, silt and clay fractions. The organic content of the samples was assessed using an ashing technique and porosity was assessed using the clod method. The organic content of the samples used to ascertain textural characteristics was destroyed using an H_2O_2 bath, however as Mikuta et al., (2005) note, this is not always a reliable methodology for ascertaining organic content of a sample, thus it should be assumed that the ashing technique provides a more representative measure.

Soil cores roughly 16.5 cm in diameter, encompassing the top 30 cm of the soil profile were also extracted at each site. This allowed for a single sample, consisting of much of the active rooting depth, to assess the K_s within this layer as a single unit as opposed to a large number of sub-samples. The samples were acquired using a technique similar to those by which porosity samples are gathered, with an outer tubing serving as a shield for the inner core, allowing for minimal disturbance within the sample. The saturated hydraulic conductivity of several samples was assessed in the laboratory using a constant head method (Black et al., 1965). Saturated soil moisture content was assessed as well, with saturated soil moisture content assumed to be 93% of the porosity of the samples (Williams et al., 1992; Minasny et al., 1999).

3.3.2.2 Numerical Analysis

The soil analyses provided parameters which can be used in a number of pedotransfer functions, relating commonly measured soil properties to the hydraulic parameters of a soil matrix. The soil properties were evaluated against the pedotransfer functions derived from the European HYPRES database (Nemes et al., 2001) and those derived from the UNSODA database (Rawls et al., 1983b; Saxton et al., 1986). The pedotransfer functions derived from the UNSODA data were used because they are primarily based on North American soils. The pedotransfer functions derived from Rawls et al., (1983b) were used to estimate the bubbling pressure (ψ_b) [L] and the pore-size distribution (λ) (Vieux, 2004):

$$\begin{aligned}\psi_b = \exp[&5.3396738 + 0.1845038(C) - 2.48394546(\phi) - 0.00213853(C)^2 \\ &- 0.04356349(S)(\phi) - 0.61745089(C)(\phi) + 0.00143598(S)^2(\phi)^2 \\ &- 0.00855375(C)^2(\phi)^2 - 0.0001282(S)^2(C) + 0.00895359(C)^2(\phi) \\ &- 0.00072472(S)^2(\phi) + 0.0000054(C)^2(S) + 0.50028060(\phi)^2(C)]\end{aligned} \tag{3.2}$$

$$\begin{aligned}\lambda = \exp[&-0.7842831 + 0.0177544(S) - 1.062498(\phi) - 0.00005304(S)^2 \\ &- 0.00273493(C)^2 + 1.11134946(\phi)^2 - 0.03088295(S)(\phi) \\ &+ 0.00026587(S)^2(\phi)^2 - 0.00610522(C)^2(\phi)^2 - 0.0000235(S)^2(C) \\ &+ 0.00798746(C)^2(\phi) - 0.00674491(\phi)^2(C)]\end{aligned} \tag{3.3}$$

where S is the sand content (%), C is the clay content (%) and ϕ is the total porosity. The wetting front suction (ψ_{wf}) [L] was determined using the Brooks – Corey relations (Brakensiek, 1977; Rawls, 1983b; Chahinian et al., 2005):

$$\psi_{wf} = \frac{2+3\lambda}{1+3\lambda}\frac{\psi_b}{2} \tag{3.4}$$

The saturated hydraulic conductivity of the soils of each nano-catchment was then determined via (Rawls et al., 1983a):

$$K_s = \alpha * \frac{(\theta_s \psi_b)^2 (\lambda)^2}{(\lambda + 1)(\lambda + 2)} \tag{3.5}$$

where α is a constant accounting for gravity and a variety of fluid constants (21 cm^3/sec), and θ_s is the saturated soil moisture content (0.93 * ϕ).

During field activities rainfall did not exceed the hydraulic conductivity, so infiltration could not be directly estimated. Instead the depth of water infiltrating to saturation was modelled to examine fine-scale spatial variation as a result of both antecedent moisture and spatial differences in K_s. Infiltration up to saturation (F_s) was estimated using the Mein and Larson (1973) model for preponded infiltration under conditions where the rate of application exceeds the saturated hydraulic conductivity of the soil matrix:

$$F_s = \frac{\psi_{wf} \cdot \Delta\theta}{i / K_s - 1} \tag{3.6}$$

where F_s [L] is the cumulative infiltration up until ponding, $\Delta\theta$ [L/L] is the soil moisture deficit (θ_s-θ_i), θ_s is the saturated soil moisture content and θ_i is the initial soil moisture content, r [L/T] is the intensity of the precipitation input and ψ_{wf} is the pressure at the wetting front and is taken to be equal to the ψ_{av} (average wetting front suction) (Moore et al., 1980). The use of this model provides information on the variability of infiltration up until ponding across the site. It also provides sufficient information to compare volumetric estimates of infiltration prior to saturation across the site at several measurement scales. Estimates of initial soil moisture at Glen Haffey totaled 160 points, while at Crawford Lake 173 points were used. Thiessen polygons were then constructed around each soil moisture point, and the depth equivalent of water infiltrating prior to saturation was then estimated using the measurements for each polygon. These depths were then converted to volumes per polygonal unit to compare against the volumetric estimates for the whole site. This use of multiple point measures to assess soil moisture allows for the bias that would normally occur from using a single point measure of initial soil moisture content to be minimized. Estimates at this scale were derived throughout the site via Thiessen polygons to analyze the variability of infiltration as a result of soil moisture variability. The volumetric infiltration to saturation for the entire bowl was then estimated through the use of estimates of K_s as derived from Rawls et al., (1983a), constant head permeameter, and through an average value of K_s derived from the infiltrometer experiment.

3.3.2.3 Terrain Indices

The terrain indices derived at each site include the local slope, wetness index (Beven and Kirkby, 1979), profile curvature and planiform curvature. All of these terrain indices can influence topographically-driven movement of water. Planiform curvature describes flow convergence and divergence, while profile curvature describes gravity-driven surficial acceleration of water downslope. These indices are all interrelated, having non-independent effects on the movement of water, assuming topographically-driven redistribution.

Each terrain measure was derived from a digital elevation model at resolutions of 0.25 m and 0.34 m for Glen Haffey and Crawford Lake, respectively. The measures were then averaged over a 5*5 cell moving window (i.e. within ~ 0.63 m at Glen Haffey and

~0.85 m at Crawford Lake) to better match the mean conditions in close proximity to the tensiometers. This technique was employed because, while tensiometers measure matric potential at a point, they are influenced by the conditions surrounding that measured point, and by integrating data from around these points we then account for this influence on the point of measurement. This also minimizes the effect of small elevation variation in the dense survey data having an exaggerated impact on these topographic measures.

3.4 Results

3.4.1 Soil Properties

The range of textural properties at Glen Haffey in the near-surface was fairly limited, all falling within the sandy loam classification over the study site (Table 3.1). For that reason average textural attributes were used throughout the bowl (Table 3.2) to estimate the hydraulic properties over the site. The texture at Crawford Lake is a silty loam at the surface. The soil at the subsurface (15 cm) is composed of a larger sand content (sandy loam) (Andrew-McBride and Graniero, in review), than that measured here indicating past sorting. However, because of the spatial persistence and reduced K_s that is typical of a silty loam layer, as a compared to sandy loam, this textural data was used to determine the hydraulic properties via pedotransfer function within the nano-catchment.

Since soil texture and organic matter varied little throughout the sites, it was assumed that bubbling pressure and the pore-size distribution index were roughly uniform at each site (Table 3.3). It should be noted though that some deviation is to be expected as a result of small changes in textural properties at these locations that were not uncovered in this study. A bubbling pressure for the Glen Haffey was determined to be 11.71 cm, similar to those found by both Panian (1987) and Carsel and Parrish (1988) at 9.01 and 13.33 cm, respectively. The bubbling pressure of Crawford Lake, 23.8 cm, is less the documented bubbling pressures noted by those authors for similarly textured soils. This is not surprising because of the relatively high amount of sand (33%) found in this silty loam unit. Using the Brooks and Corey (1964) equation (Equation 3.4), the wetting front suction was determined to be 8.61 cm at Glen Haffey and 17.80 cm at Crawford Lake. Using these hydraulic properties the saturated hydraulic conductivity was determined to be 12.34 cm/hr and 3.04 cm/hr for Glen Haffey and Crawford Lake respectively, as determined using pedotransfer function (Rawls et al., 1983a).

Table 3.1 Soil textural properties at three surface sites within the Glen Haffey (GH) and Crawford Lake (CL) nano-catchments

-	GH-S1	GH-S2	GH-S3	Average Values	CL-S1	CL-S2	CL-S3	Average Values
Sample (g)	10.00	10.01	10.00	-	10.00	10.00	10.01	-
Organic Content (%)*	2.14	2.60	1.90	2.21 (0.70)	2.59	2.20	3.10	2.63 (0.90)
Sand (%)	57.37	62.82	57.27	59.15 (5.55)	31.70	36.93	30.76	33.12 (6.17)
Silt (%)	31.68	25.23	33.02	29.97 (7.79)	52.94	50.89	52.50	52.10 (2.05)
Clay (%)	10.95	11.94	9.70	10.86 (2.24)	15.36	12.19	16.75	14.76 (4.45)

* Derived from the use of H_2O_2

Table 3.2 Average nano-catchment soil properties for each site

Site	Texture	Sand (%)	Silt (%)	Clay (%)	Oraganic Matter (%)	Organic Content (58% OM)	Porosity (v/v)	Saturated Soil Moisture θs (%)	Bulk Density (g/cm3)
Glen Haffey	Sandy Loam	59.15	29.98	10.87	3.76	2.18	0.48	44.62	1.27
Crawford Lake	Silty Loam	33.13	52.11	14.76	3.38	1.96	0.52	48.27	1.38

Table 3.3 Average nano-catchment soil hydraulic properties for each site

Site	Bubbling Pressure (cm)	Wetting Front Suction (cm)	Pore Size Distribution	K_s (Rawls et al., 1983a) (cm/hr)	K_s (permeameter) (cm/hr)	Average Field K_s (cm/hr)
Glen Haffey	11.71	8.61	0.3758	12.54	18.25	22.10
Crawford Lake	23.83	17.80	0.3414	3.04	6.64	18.83

3.4.2 Infiltration and Saturated Hydraulic Conductivity

While monitoring the infiltrometers, it was evident that considerable spatial variability exists within both nano-catchments. The infiltrometer locations span 4.5 vertical meters at Glen Haffey and 3.5 m at Crawford Lake, and within this limited vertical gradient infiltration rates vary considerably (Table 3.4). The GH-I4 infiltrometer for example has a local slope of 21.5% and a mean infiltration rate of 63.3 cm/hr under a constant head of 5 cm. This infiltrometer was installed, as noted, to evaluate the effect of placing an infiltrometer on a gradient; this effect is evident in that the rate nearly doubles the rate measured at any other point. This 'error' is especially important where such measurements are used to evaluate infiltration on an areal basis (i.e. catchment hydrologic models).

Infiltration rates at the bottom of the nano-catchment are higher than at other locations within the bowl. The infiltration rate of 33.2 cm/hr at GH-I2 is heavily influenced by macropores, which are visible at the surface in the area surrounding the infiltrometer. The sites at the higher relative elevations are fairly similar to one another (20.8 and 25.2 cm/hour, respectively) even though they are at opposite margins of the nano-catchment. In determining approximate values of K_s, site GH-I4 was excluded from further hydrological analysis because of the effect topographic gradient has on redistributing the infiltrating water laterally below the infiltrometer ring. Even with the exclusion of site GH-I4, K_s deviated considerably as determined by the infiltrometers (Table 3.4).

The Crawford Lake pattern of infiltration, like Glen Haffey, is spatially variable. The measures of infiltration from both November and June were integrated in order to facilitate comparison. In discussing site CL-I1 and site CL-I32, we find a higher rate of infiltration, at 34.39 cm/hr and 35.30 cm/hr, respectively (Note: double digit infiltrometer IDs are indicative of June measurements). This high rate of infiltration can be assumed to be a result of root activity within this area of native vegetation. This observation also corroborates the use of both collection periods as a single sample, as the two sites are in very close proximity and it is this area which would be most affected by changes caused by root growth and decay between growing seasons. Site CL-I5 (35.94 cm/hr) is situated within a zone where the fallow agricultural area transitions into native grasses. Thus large scale macropores in the subsurface are present from previous field tillage practices and current root activity. Significantly lower measured rates of infiltration occurred at sites CL-I2, I3 and I4 (3.95 – 6.34 cm/hr), showing an absence and/or a reduction in preferential flow paths within the soil matrix at these points. The considerable differences in the rates at the proximal sites CL-I5 and I4 (8.2 m apart) and the similar rates at sites CL-I42 and I32 (21.3 m apart) demonstrates the fact that although the bulk soil properties are relatively uniform, the spatial variability of infiltration is high and that the parameters associated with the infiltration of water are also spatially variable. This suggests that estimates of the amount of water that may be infiltrated over the entire nano-catchment can be heavily influenced by the spatial density of input measurements used in the estimate. Had a single measure of infiltration been made at either Crawford Lake or Glen Haffey, much of the fine-scale variability would have been lost and considerable differences in the partitioning of pre-ponded infiltration is likely. Again, this is critical in the case where point measures of infiltration are used in conjunction with local groundwater recharge models.

Table 3.4 Infiltration and estimated hydraulic conductivity for Glen Haffey (GH) and Crawford Lake

Infiltrometer	Average Infiltration Rate (cm/hr)	Total Cumulative Infiltrated Water (cm)	θ_e (%)	$1/K_s$	K_s (cm/hr)
GH-I1	**20.82**	100.841	0.270	0.049	**20.352**
GH-I2	**33.22**	156.863	0.226	0.030	**32.813**
GH-I3	**11.87**	50.166	0.210	0.087	**11.460**
GH-I4	**63.32**	246.245	-	-	-
GH-I5	**25.23**	65.190	0.214	0.041	**24.540**

Infiltrometer	Average Infiltration Rate (cm/hr)	Total Cumulative Infiltrated Water (cm)	θ_e (%)	$1/K_s$	K_s (cm/hr)
CL-I1	**34.39**	188.949	0.214	0.029	**34.174**
CL-I2	**3.95**	21.136	0.185	0.266	**3.763**
CL-I3	**6.35**	32.086	0.231	0.164	**6.102**
CL-I4	**4.08**	40.234	0.182	0.251	**3.981**
CL-I5	**35.95**	210.848	0.185	0.028	**35.772**
CL-I22	**19.24**	116.119	0.275	0.053	**18.990**
CL-I32	**35.30**	132.672	0.258	0.029	**34.922**
CL-I42	**13.75**	56.532	0.237	0.074	**13.434**

3.4.3 Volumetric Infiltration to Saturation under Hypothetical Rainfall Inputs

Infiltration to saturation (F_s) was determined through the use of the initial soil moisture values collected April 19, 2005 during a relatively dry period at Glen Haffey. The soil moisture measurements made June 15 and 16, 2005 were used to create the initial Crawford Lake soil moisture surface as noted. A resultant distribution of soil moisture was constructed using Thiessen polygons to relate points of known soil moisture to the areas closest to these points (Fig. 3.4). These areal units were then matched with the most proximal measure of saturated hydraulic conductivity as determined by infiltrometer. Therefore, F_s can be calculated (Mein and Larson, 1973) (Equation 3.3) for each polygonal unit at different rainfall intensities. Differing areas within the bowl shifted from infinite infiltration prior to saturation at rainfall rates of 20 cm/hr, to very limited infiltration prior to saturation as limited by K_s at 60cm/hr (Table 3.5). The time to ponding at each site of known saturated hydraulic conductivity can be assumed to be equal to K_s / i, where i is the rainfall intensity (Mein and Larson, 1973), thus at low rates of precipitation, total surface saturation or ponding does not occur. The total area-weighted volumetric infiltration prior to saturation for the aggregated subunits of the nano-catchment are compared to the total volumetric infiltration prior to saturation within the 'averaged' bowl (Table 3.5). Under lower intensity events the amount of water able to penetrate the soil profile is dramatically higher where the spatial distribution of K_s is integrated into the analysis. Where i is less than K_s non-ponded conditions go on indefinitely since the rate of infiltration is equal to precipitation. In the overall bowl, estimates of pre-ponded infiltration by the constant head permeameter and Rawls et al., (1983a) are substantially less than that determined by field methods in either the polygonal units or through the averaged K_s from the infiltrometers. This is a result firstly of the fact that Rawls does not integrate macropore flow in characterizing K_s, and secondly that the permeameter measure of K_s at both Glen Haffey and Crawford Lake were substantially less than some of the points determined via infiltrometer measures.

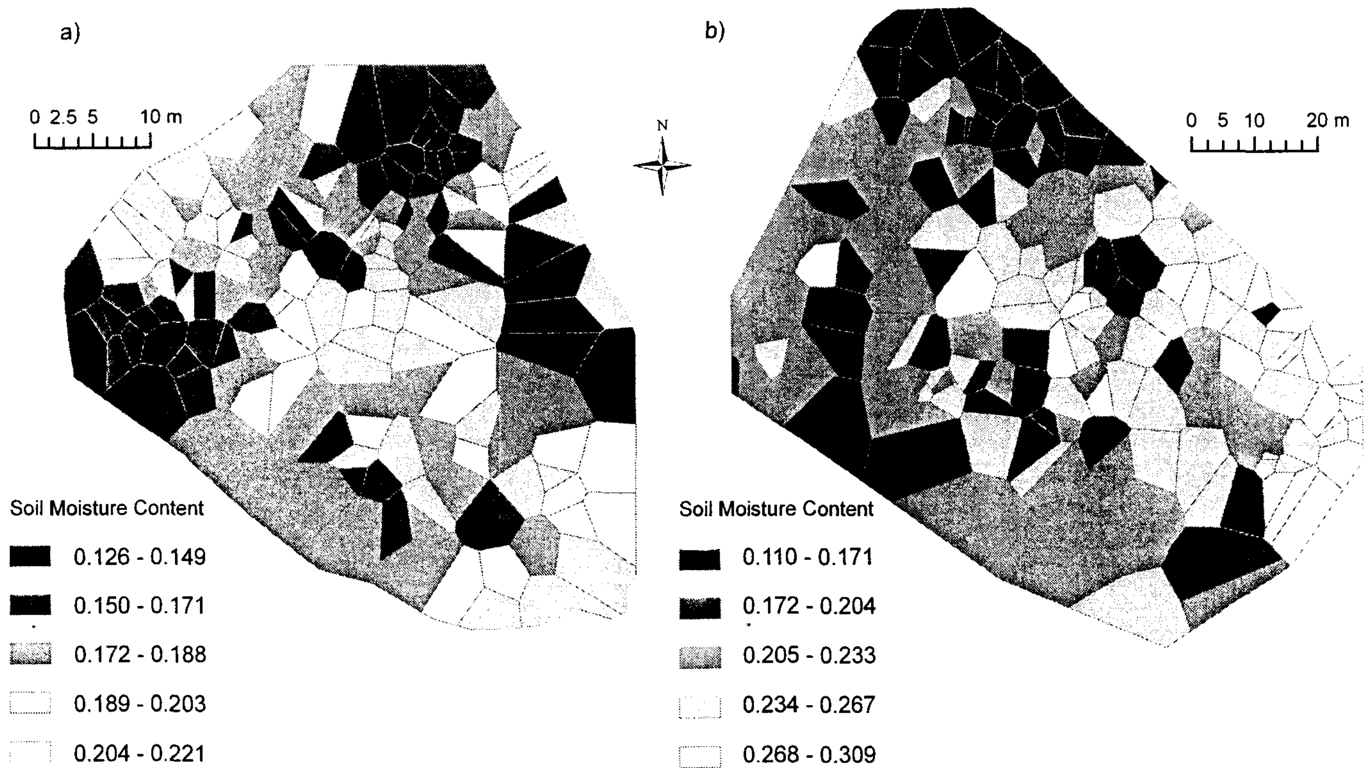


Figure 3.4 Initial Soil Moisture Conditions at a) Glen Haffey as derived from April 19, 2005 data acquisition; and b) Crawford Lake from June 15 and 16, 2005 acquisition.

Table 3.5 Volumetric (m^3) estimate of infiltration to saturation for each nano-catchment under varying intensities of rainfall as derived from four methods of estimating saturated hydraulic conductivity

Glen Haffey Technique of Determining K_s	Volumetric Infiltration (m^3) prior to saturation at: 20 cm/hr	30 cm/hr	40 cm/hr	50 cm/hr	60 cm/hr
Averaged from Infiltrometers	Infinite*	110.84	51.88	33.87	25.14
Constant Head Permeameter (m^3)	192.86	53.63	31.14	21.94	16.94
Rawls et al. (1983a)	77.34	33.04	21.01	15.40	12.16
Aggregeted Polygons and Proximal Infiltrometer	Infinite**	Infinite**	65.90	34.68	23.96

Crawford Lake Technique of Determining K_s	Volumetric Infiltration (m^3) prior to saturation at: 20 cm/hr	30 cm/hr	40 cm/hr	50 cm/hr	60 cm/hr
Averaged from Infiltrometers	1111.59	348.14	206.39	146.67	113.75
Constant Head Permeameter	163.26	93.36	65.37	50.30	40.87
Rawls et al. (1983a)	58.74	36.95	26.96	21.22	17.49
Aggregeted Polygons and Proximal Infiltrometer	Infinite**	Infinite**	927.81	262.53	169.77

* K_s not exceeded for the entire bowl

**K_s not exceeded for portions of the bowl

3.4.4 Matric Potential

In addition to evaluating F_s within the bowl at various scales of influence, tensiometric data collected during a rainfall event was evaluated. While the study included data spanning the period from April 19 to April 23, 2005, this first day of collection is assumed to be a period at which tensiometers were equilibrating with the surrounding matrix and therefore this data are excluded from study. However, the following four days provide a significant amount of information on the spatial variability of suction throughout the nano-catchment (Fig. 3.5 a and b). While tensiometers were inserted to 15 and 30 cm, generally the tensiometers at 30 cm depth showed very little change during precipitation events and in the immediate period following. At Glen Haffey the tensiometers inserted to 15 cm illustrate two tendencies based on their relative positions within the nano-catchment. First, those tensiometers that are found at sites of relatively lower elevation (GH-T3) are prone to react more rapidly to a rainfall event then those sites at higher relative elevation (GH-T5). Second, the wetting front at these lower sites is less diffuse than at higher elevations, meaning the wetting front advances with a more abrupt change in suction.

In addition to the variability related explicitly to elevation, other topographic indices can also be related to matric suction. These measures at each tensiometer cluster for both Glen Haffey and Crawford Lake are shown in Table 3.6. The relationship can only be made at a superficial level, given the limited number of sites. For example, site GH-T1 and T5 both display delayed responses to precipitation inputs, and they are the locations within the Glen Haffey site that have a convex profile curvature and have lower wetness index values reflecting minimal upslope catchment area. The wetness index also is indicative of GH-T3 having less suction (i.e. moister conditions) during precipitation periods which is reflected in the data from April 20 and 23. Ideally GH-T2 would have provided additional information in terms of the wetness index, but as a result of the failure of the 15cm tensiometer at this location the relationship could not be corroborated with any other data source. GH-T6 shows that while being at a mid-elevation relative to the rest of the nano-catchment, matric suction at this point differs as the area around the tensiometer can be considered a 'sub-catchment' of the nano-catchment. This area is also influenced by the presence of leaf litter which reduces the effect of topography on the soil moisture pattern and also has an effect on suction at this point (Andrew-McBride and Graniero, in review).

At Crawford Lake little information can be derived from topographic indices in terms of the resulting measured matric suction (Fig. 3.6). One can note though that suction is greatest (i.e. more negative) at relatively higher elevations. This indicates a flow of water downhill, as tensiometer CL-T3 at 15 cm measures dramatically less suction than either CL-T1 or T2 at the same depth. However beyond that single observation no clear horizontal pattern emerges, with respect to the terrain indices. This is primarily a result of the limited dataset, which is composed of data from a single day where the hydrologic regime did not change. This lack of correlation is also tied to the fact that a minimal topographic gradient exists within the bowl, thus these measures provide less insight as compared to the more undulating Glen Haffey site. At depth, the tensiometers at 30 cm experienced less suction than those at 15 cm, for all but site CL-T3. This is indicative of an evaporation dominated environment, which would be expected given the lack of precipitation at the site.

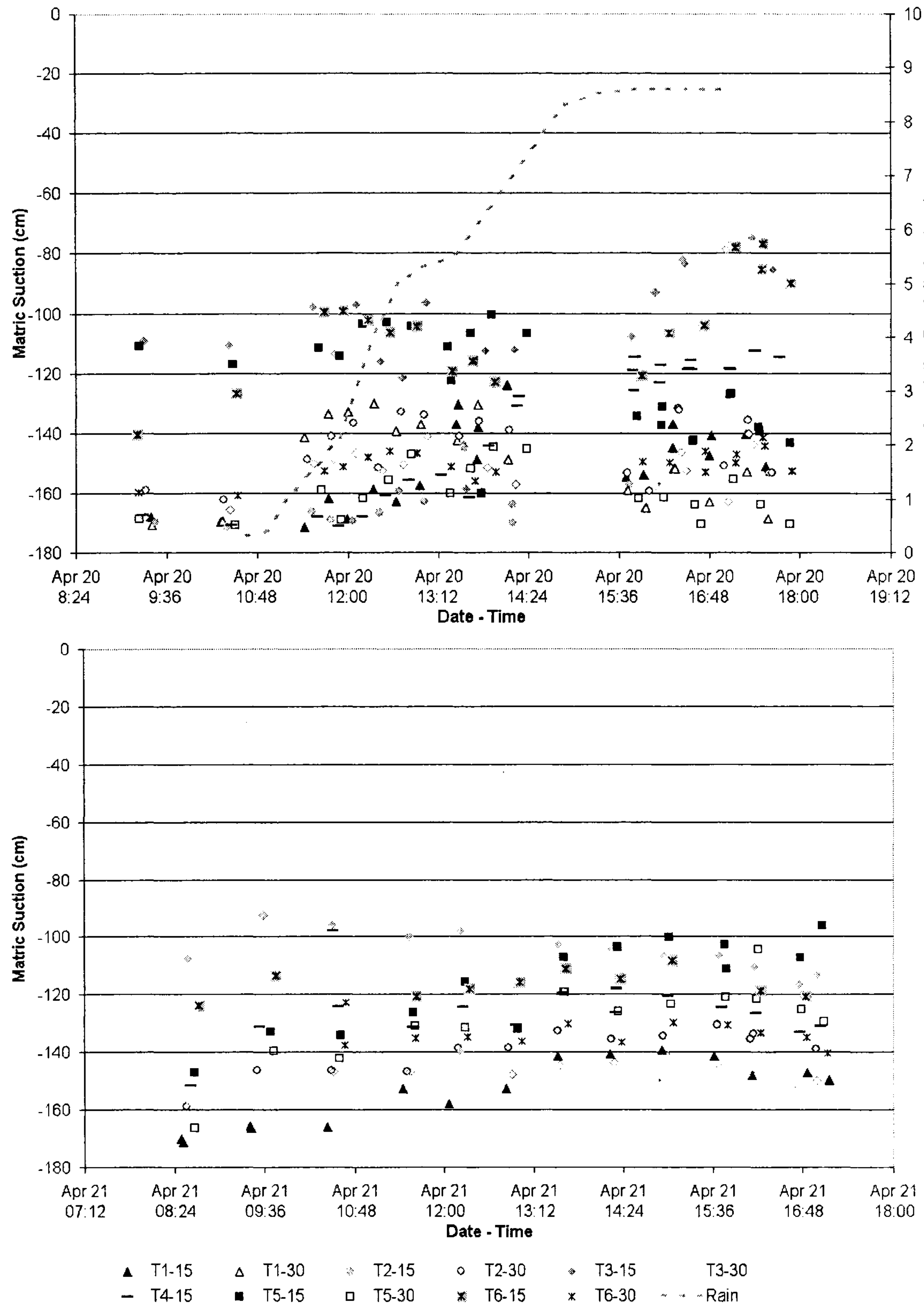
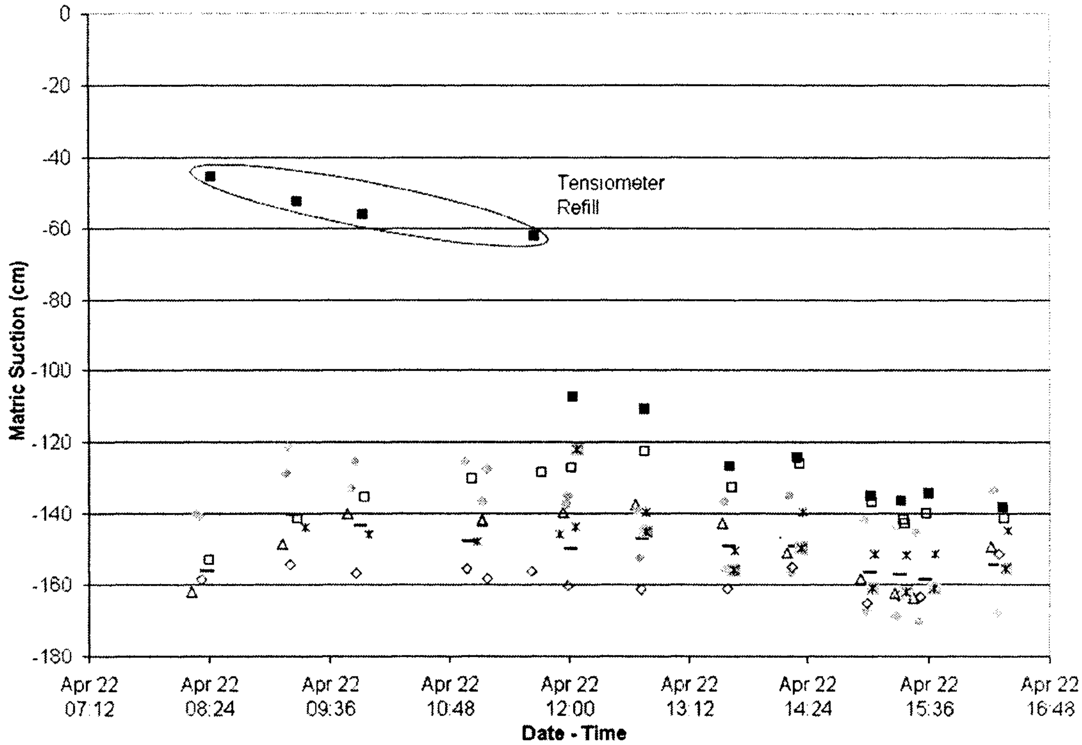


Figure 3.5a Evolution of matric suction at six sites within the Glen Haffey nano-catchment, April 20-21, 2005.



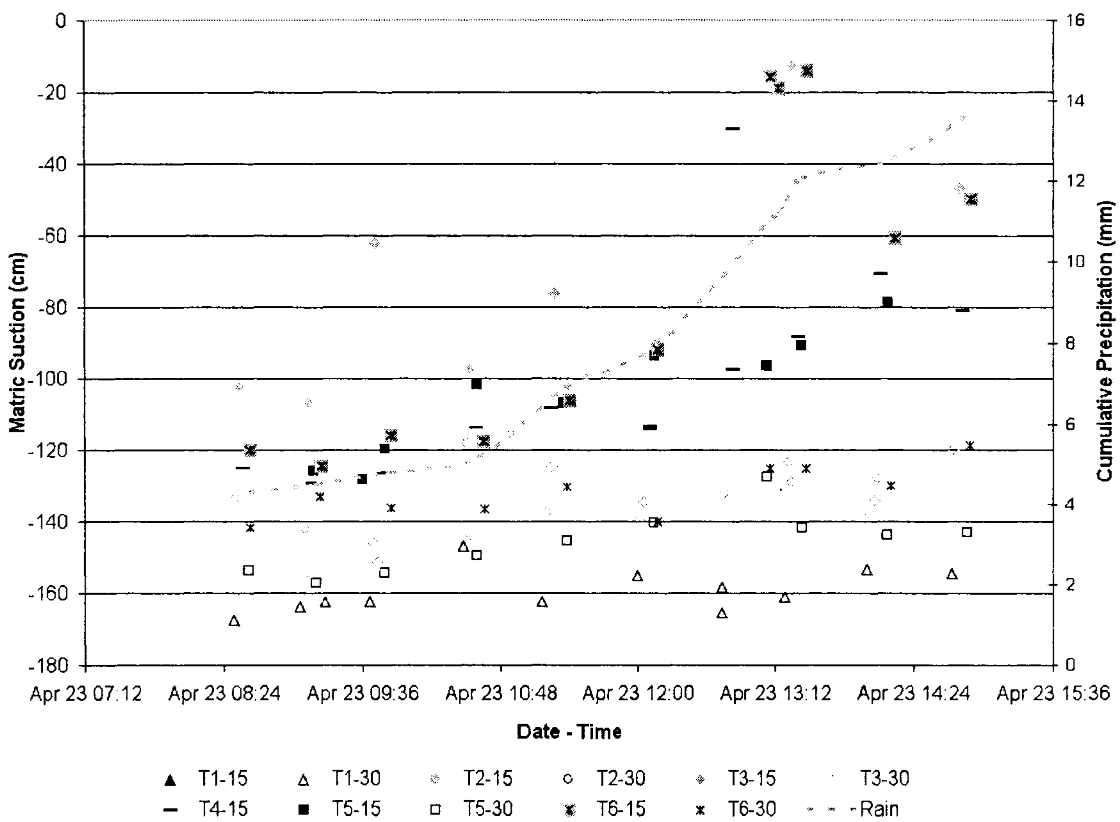


Figure 3.5b Evolution of matric suction at six sites within the Glen Haffey nano-catchment, April 22-23, 2005.

Table 3.6 Terrain attributes at each at tensiometer location within both the Glen Haffey (GH) and Crawford (CL) nano-catchments.

ID	Elevation (m)	Wetness Index	Slope (%)	Profile Curvature	Planiform Curvature
GH-T1	418.15	3.33	5.88	-4.50	1.28
GH-T2	415.72	8.87	6.10	4.19	-0.93
GH-T3	415.06	8.85	2.15	3.01	-4.82
GH-T4	416.52	7.48	1.00	3.37	-0.19
GH-T5	419.25	3.51	3.46	-4.80	5.08
GH-T6	417.38	7.25	2.76	0.81	-3.30

ID	Elevation (m)	Wetness Index	Slope (%)	Profile Curvature	Planiform Curvature
CL-T1	307.68	1.22	4.02	0.09	0.36
CL-T2	306.77	4.45	3.51	-0.50	-0.46
CL-T3	305.90	2.73	3.18	0.61	0.05
CL-T4	306.98	5.72	4.20	-1.15	0.08
CL-T5	307.02	8.52	2.43	-0.09	-0.09

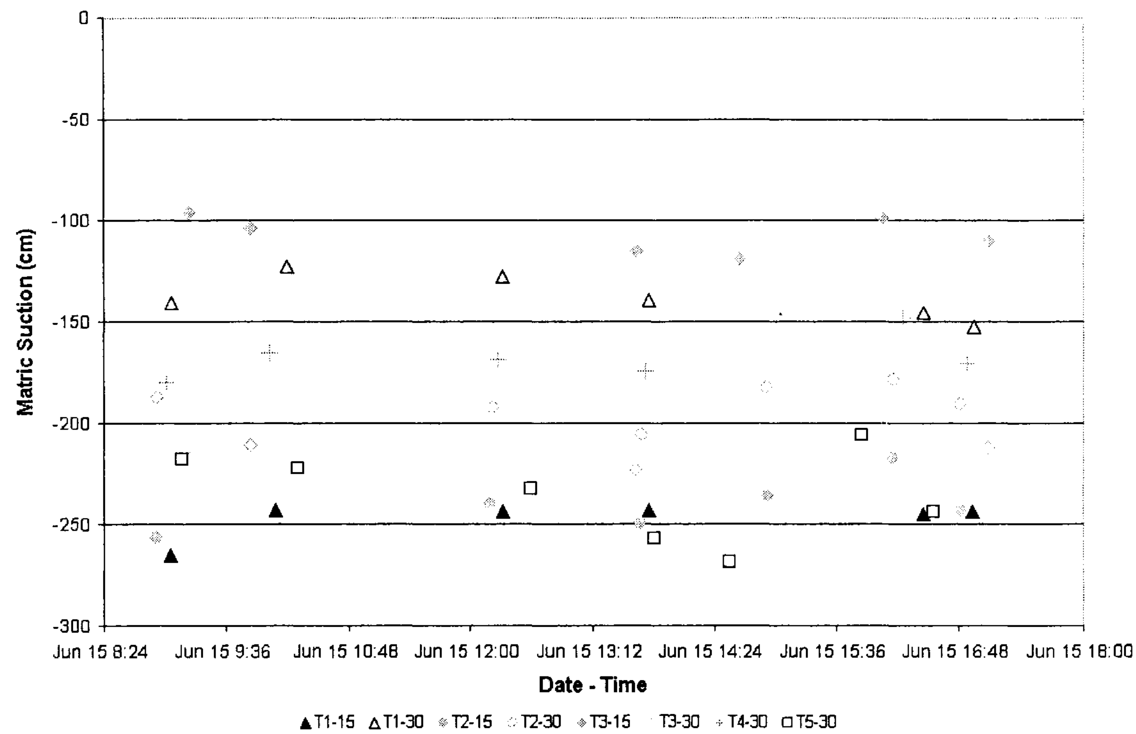


Figure 3.6 Matric suction at five sites within the Crawford Lake nano-catchment, June 15, 2005

3.5 Conclusion

Considerable spatial variability in the attributes associated with infiltration and specifically infiltration to saturation, F_s, occurs over these sub-hectare plots. These differences in the amount of water infiltrated prior to surficial saturation are important to note because under standard assessment much of the variability, and specifically the bias, between these measures would go unnoticed. It is especially important where one measure of the hydraulic characteristics of a site are used to model infiltration and overland flow. In the case of soil texture, models infer hydraulic characteristics from generalized maps with differing accuracies and scales. While we know that within these units variability exists much of this variability goes unaccounted for in catchment models as they are parameterized so that they 'work' and are fit to known output. Within catchment models accounting for this fine-scale variability is difficult, but it should be noted.

This is not to say that bias does not exist within the data presented here. The hydraulic characteristics derived from soil samples incorporate a total surface area of 14.7 cm^2 at each site, while the permeameter accounts for 132.7 cm^2 and the infiltrometers account for 192.4 cm^2 at Glen Haffey and 307.9 cm^2 at Crawford Lake. Obviously outside of these measured units variability exits, however some this variability is accounted for through the use of fine-scale soil moisture measurements. Although they only represent a few square centimetres, the sheer number of data points provides insight into the overall bowl dynamics. The use of Thiessen polygons to account for areal differences also allows for soil moisture differences to be observed and integrated into the modelling of infiltration at this scale of measurement.

The variability among the measured rates of infiltration is not surprising. It has been noted elsewhere that the range by which infiltration is spatially auto-correlated is less than 10 m (Loague and Gander, 1990; Sullivan et al., 1996). The randomness of infiltration rate is a result of the total spatial variability of *Ks* (Govindaraju et al., 2006), soil moisture (Grayson and Western, 1998; Andrew-McBride and Graniero, in review) and soil water retention (Pachepsky et al., 2001). All three of these variables vary for a number of reasons including texture, macropores, topography, rainfall, etc., which account for the semi-randomness seen in infiltration modelling. The effect of macropores is such that a soil matrix is able to conduct water away from the surface at a much greater rate than would be estimated from hydraulic conductivity. Infiltration is also obviously affected by the rate and variability of precipitation (Goodrich et al., 1995; Govindaraju et al., 2006). However, little variability in precipitation occurred over the Glen Haffey nano-catchment, minimizing the effect of this variability at this scale of study.

This variability in infiltration, as mentioned, is partially a function of the variability of matric suction over the nano-catchment. This is a factor of the wetting front not proceeding downwards uniformly throughout the site. This, like infiltration, is controlled by the interaction of soil texture, topography and macropores in general. However, given the relative uniformity of the near-surface soil texture within these two nano-catchments, topography and macropores are of more importance. This importance demonstrates the variability that can occur within typical model grid cells which are typical of catchment models such as TOPMODEL (Beven et al., 1995).

While no distinct pattern exists here, an argument can be made that at the Glen Haffey site topographic indices provides some insight into the local soil retentive properties, which counter Pachepsky et al., (2001), who found no correlation between water retention and curvature. The hydrologic dynamism of the collection period provided evidence that topographic variables have an influence on water retention during wetting conditions at Glen Haffey. The relationship between curvature and water retention properties is not apparent at Crawford Lake. However, the topographic gradient at Crawford Lake is much smoother which limits curvature and is similar to the conclusions Pachepsky et al., (2001) made at their gently sloping site.

While the goal of this study was to monitor and account for the variability of hydraulic gradients associated with infiltration, the most important observation is that numerous biases can be introduced in using point estimates of any of these measures. The estimates of attributes associated with pedotransfer functions, while being critical in modelling must be used with caution because of the inability of these tools to incorporate all variability. Where data exists this variability should be investigated and accounted for in modelling. The use of single point data to investigate spatially variable phenomena should also integrate some measure of error, as even within a single soil unit much variability exists. The variability investigated here can be seen to be a result of topography, macropores and to a degree soil moisture and vegetation differences.

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CHAPTER IV – Conclusion

4. 1 Discussion and Conclusion

The spatial and temporal variability of fine-scale hydrologic structures was explored in this study and it is clear that a number of parameters affect these structures at various scales. Topography within the nano-catchments influenced hydraulic conductivity, soil moisture, matric potential and infiltration to saturation, but the variability cannot be explained solely based on this physical attribute. Many fine-scale features modify the topographic control on the hydrologic parameters, including the absence / presence of vegetation, the effect of leaf litter, soil texture, macropores, and the precipitation regime. While all of these structures clearly affect the spatial pattern of soil moisture, and matric potential in particular, quantifying the role of each structure is difficult.

The timing of precipitation has a role in governing the spatial pattern of soil moisture. The spatial pattern of soil moisture under 'wet' conditions is spatially auto-correlated over greater distance than under 'dry' conditions. This is a result of the homogenizing effect of rainfall on the overall landscape. This conclusion is substantiated by the fact that correlation lengths (i.e. range) increase during moister conditions (19.6 m) from that of dry conditions (14.6 m) at Glen Haffey. The spatial density of the sampling scheme also has an effect as soil moisture reaches a correlation length of 44 m at Crawford Lake, where the sample density is considerably lower. This range is typical of other studies that have looked at the auto-correlation of soil moisture at sparser sampling densities (Grayson and Western, 1998; Lookingbill and Urban, 2004). While rainfall provides this homogenizing effect, the underlying pattern of soil moisture is affected by other physical nano-catchment attributes. These attributes include both topography and vegetation, in both the form of active photosynthesizing vegetation and leaf detritus. At Crawford Lake the pattern of soil moisture in relation to topography is enhanced by the presence of active vegetation, whereas at Glen Haffey leaf litter serves to reduce what relationship there is between soil moisture and elevation. The variation in vegetative land cover heightens the variability of soil moisture throughout both sites.

Spatial auto-correlation within the soil moisture pattern has an effect on the other hydrologic processes that occur within each nano-catchment. Infiltrometer tests display variability in infiltration rates on the scale of meters within each bowl, and since infiltration is tied directly to antecedent soil moisture, one can infer that infiltration auto-correlation is less than 14.6 m at Glen Haffey and less than 44 m at Crawford Lake. However, the auto-correlation is likely much less than either of these two measures as a result of the spatial heterogeneity of macropores and, to a lesser degree, vegetation within each nano-catchment. This observation is similar to conclusions drawn by Loague and Gander (1990) who found spatial-autocorrelation to exist at ranges of less than 10 m. This spatial variability is also displayed in measurements of matric suction that vary with both horizontal distance and vertical distance within the soil matrix. Differences in suction between depths of 15 cm and 30 cm within both nano-catchments are prevalent. This deviation between the two depths is reflective of the local hydrologic regime. Under conditions that include or immediately follow events, suction is less in the near surface (15 cm) as compared to that at depth (30 cm). However, at drier times when soil water has both infiltrated to depth under the force of gravity and the influence of suction, and has been drawn from the near-surface through evaporative processes, suction at 30 cm is less than at 15 cm. Matric suction at 30 cm also displays less temporal variability over short timeframes (days) than that at 15 cm. Again this is a result of more prevalent evaporative forcing in the near-surface and more rapid response to precipitation. These differences highlight the fact that differing spatial and temporal processes are interactively working to affect matric suction.

The effect of topography as represented by indices derived from digital elevation models indicate some topographically driven control on the spatial measures of matric potential at the Glen Haffey site. These indices, though, do not provide conclusive evidence that matric potential over the entire site is a result of the relative topographic position of each tensiometer. The effect of topographic convergence on matric potential, like soil moisture, is only really evident in those sites where topographic convergence is extreme. This is also likely a reason why topographic indices provide little relative benefit in analyzing matric suction at Crawford Lake as the general topographic gradient within the bowl was not significant, as compared to Glen Haffey. The effect of surface curvature on matric potential shows some evidence of control at Glen Haffey, but by no means is this conclusive and at Crawford Lake little effective control can be found from curvature. The use of terrain indices at the nano-catchment scale, while producing some interesting correlations or absence thereof, cannot be conclusively stated to be the primary control of matric potential at a point. At Glen Haffey where sufficient topographic gradients exist the use of topographic indices provide greater insight into soil water dynamics. At Crawford Lake terrain indices provide considerably less insight, similar to the conclusions of Pachepsky et al., (2001) who found that in gently sloping areas terrain indices showed little utility for elucidating the spatial variability of matric suction. However at both sites topography does have an effect on the longer-term subsurface lateral redistribution of water within the bowls, which is evident in areas of sufficient topographic convergence. Other attributes interact with topography within areas of similarly textured soils, producing the patterns of soil moisture, hydraulic conductivity and matric suction.

Macropores in general lead to the greatest temporal and spatial variability in hydraulic conductivity, and consequently matric suction at a point. However as displayed throughout the study capturing this effect over space is difficult, if not impossible, thus in further study the development of proxies for this variability will be key. In this work we studied both vegetation and macropores as distinct features, which is the case in agricultural environments, however in 'natural' landscapes changes in vegetation can also be an indicator for changes in macropore presence.

The use of multiple measurements of soil moisture and hydraulic conductivity allowed for the study of infiltration across each bowl. These measures integrated both single point data from infiltrometers, along with measures of hydraulic conductivity from pedotransfer functions and permeameters, thereby allowing for a volumetric infiltration budget over the site to be calculated. The use of single measures to model infiltration across each bowl determined a substantially less amount of water infiltrating prior to surficial saturation than did the measures that integrated both the variability of hydraulic conductivity and soil moisture across the nano-catchments. This difference highlights the fact that single point measures can produce dramatically different results than those that integrate more spatial variability.

The spatial and temporal heterogeneity of the hydrologic attributes within these sub-hectare 'nano-catchments' demonstrates the potential for a great deal of bias in normal catchment models. In general, catchment models only incorporate meso-scale variability in terms of the spatial variability of land cover and topography. However, here we noted the bias that results from point measures of infiltration across an area where complex interactions alter the hydrologic regime. These point measures include those from permeameters used in determining hydraulic conductivity, single infiltrometers and those derived through the use of pedotransfer functions that incorporate only soil attributes in order to determine hydraulic characteristics of the matrix under study. The use of pedotransfer functions, while mandated in most small scale studies, induces bias by assuming that only soil properties influence these hydraulic characteristics tied to infiltration. However, we see here that a variety of nano-catchment characteristics influence the properties of a given point and specifically macropore presence, which is not mapped at even the finest scale.

The use of single point measures to model the soil moisture, hydraulic conductivity and matric suction across a field-scale study would fail to properly capture the variability that can have a dramatic effect on the infiltration regime. With increasingly smaller scale studies this bias towards single points representing larger and larger areal units homogenizes much of the intrinsic hydrologic variability found within each spatial unit. The effect of this homogenization depends on the study's objective. Where single points only measure soil moisture some of this variability can be lost without dramatic effects, however where single points monitor infiltration this variability can lead to substantial differences. In this case, not only is soil moisture monitored, but also hydraulic conductivity and matric suction, all three being affected by a number of physical properties, including both soil properties, topographic shape and effects of vegetation / macropores.

4.2 Potential changes to aid in future study

With all research there are things that one would wish to do differently if the process were undertaken again. Here, I will briefly state a few changes to this study that would have aided the analysis of the data.

One hindering factor in the analysis of this data was that both temporal and spatial variability were assessed concurrently. Although this can provide a great deal of insight into the processes, the manner in which the data is collected must fit both goals. In monitoring soil moisture through dry, intermediate and wet conditions, determining the temporal dynamism of the pattern as a result of rainfall was a major goal. However, because of the dynamic nature of soil moisture one clear pattern of data collection did not present itself. A random sample presents the benefit of not having any predetermined bias, but it also brings in the possibility that some areas of the study area may be under-sampled. A gridded or uniform pattern of data collection minimizes under sampling over the entire bowl, though it biases the data in that only specific sites are studied and any variability that exists between these points is lost. Only a few separation distances are represented, making variogram construction difficult. It also tends to lead to a 'pock-marked' surface when looking at the spatial variability through either linear interpolation or kriging. Thus I chose in this research to use a stratified random sampling technique to capture soil moisture throughout the sampling period. This sampling procedure would in general provide the best of both worlds, in accounting for the variability within points, while also not having any area within the bowl under sampled.

However, given that this was the first exercise using this type of rapid data acquisition technique a number of unknowns were present during data collection. The two most important of these are very much intertwined. The first issue was that the length of time to acquire each sample was unknown, leading to an initial sampling procedure that included too many discrete representative areal units in the sampling strategy. The relatively short period in which data was collected at both sites presented an enormous challenge in determining the correct size of these units. The units themselves also were not manually identified on the landscape thus some duplicate samples were taken in some areas, while others were missed entirely. The second issue was that within each bowl certain areas were virtually impenetrable to the insertion of the time domain reflectrometers. This was especially evident for the 30 cm samples that were not inserted in certain areas of the bowls or were only inserted in those difficult areas where they could meet the minimum soil depth. This leads to both an undersampling of these units, and to some bias within the units. Along with the need to capture soil moisture in proximity to the tensiometers, this tended to cluster the data points, most especially at Glen Haffey.

At the time of collection the variability of the soil moisture regime at both 15 and 30 cm was unknown, therefore determining representative areas by which this units could be drawn was difficult, if not impossible. The reason for this again comes back to the fact that the spatial extent of the study differed from other published sources, thus discriminating these units was left to 'best guess'. The temporal dynamism was also difficult to assess because no two points matched up exactly in terms of their x, y coordinates. Thus because of the spatial variability discussed throughout the paper matching any two points on a temporal scale was difficult.

This highlights the need for continued research into proper sampling strategies with these kinds of field tools to improve data collection procedures. The use of a nested-sampling strategy provides some definite possibilities into this problem, however this size of each nested unit again is based on the knowledge of the degree of variability over the unit of study for any phenomenon being studied.

In addition to this general issue with the sampling strategy, I would have ideally collected more soil samples in order to account for even the minimal variability that I assume to exist over the relatively small units. With only three surficial cores and 5 subsurface cores at each site the possibility that more variability exists within each unit does present itself. Again at the time of data collection, some uncertainty to the degree of variability that would be expected was unknown, however unlike soil moisture studies, this site variability could have been ascertained prior to collection. It was assumed at the time that a characteristic moisture curve could be better developed, however because of the vertical insertion of each TDR and the point measure of the tensiometers a reliable curve could not be produced, thus the need for soil samples to drive pedotransfer functions became more prominent. So like all field investigations certain things could have been done differently, however the goal of the study was sufficiently met.

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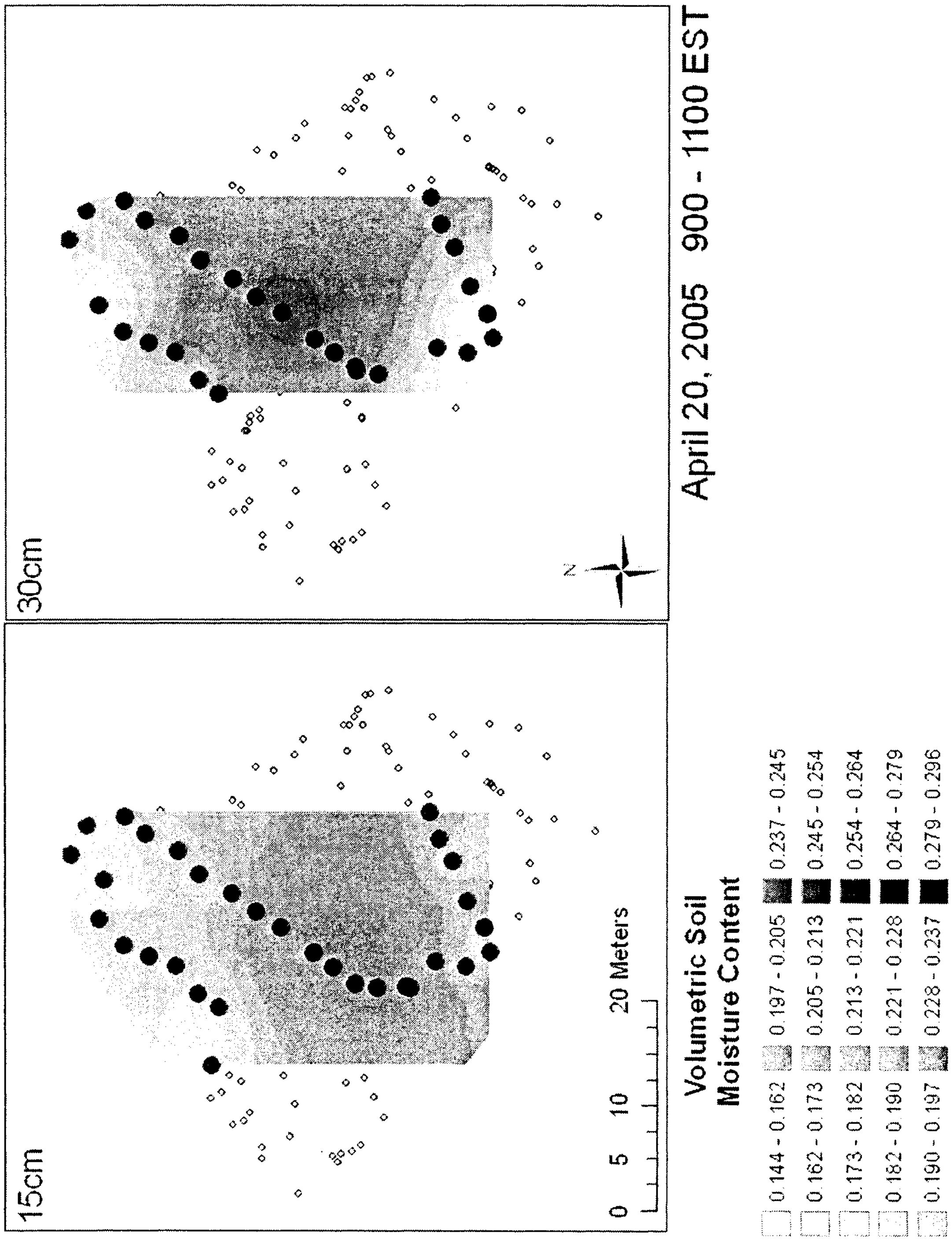
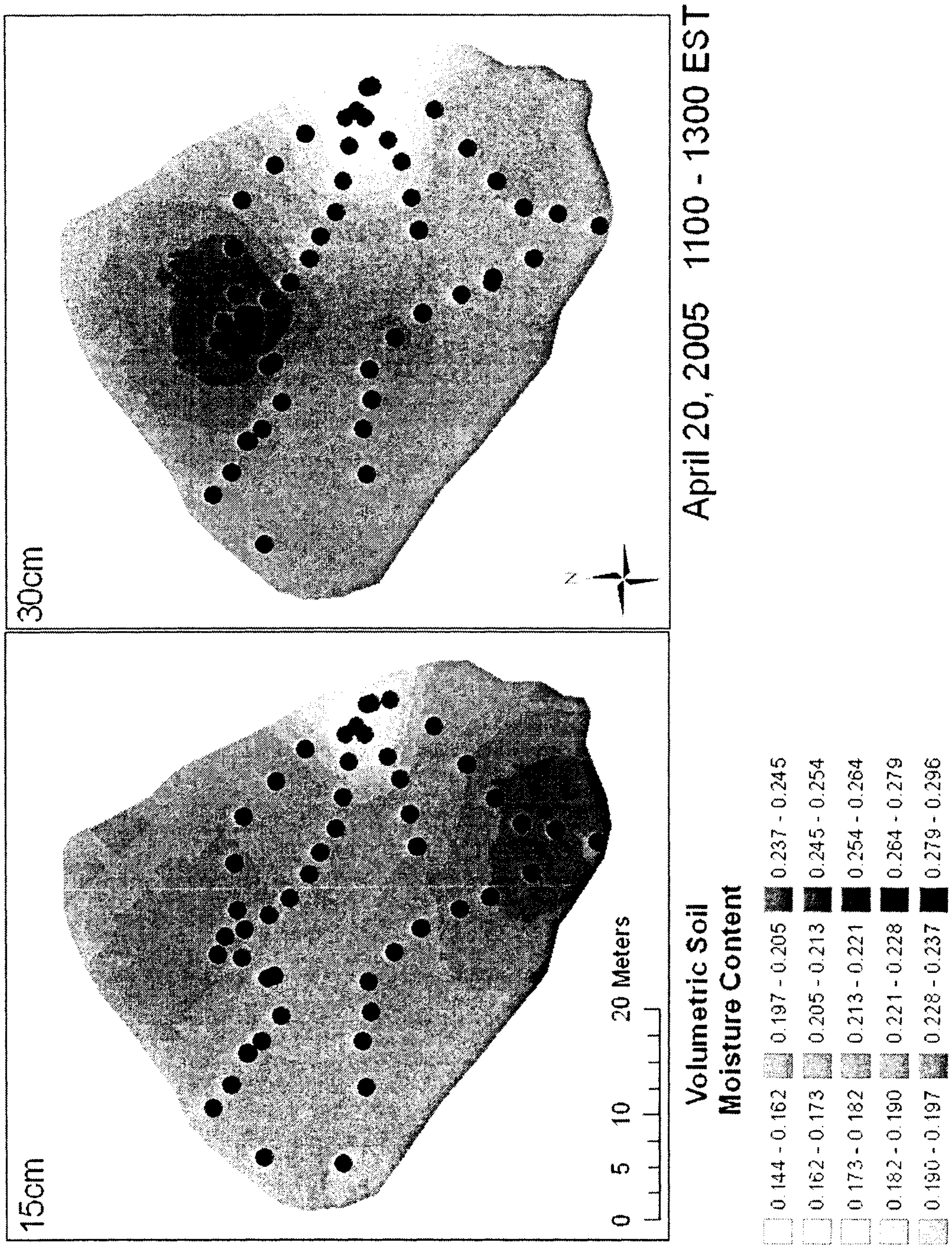
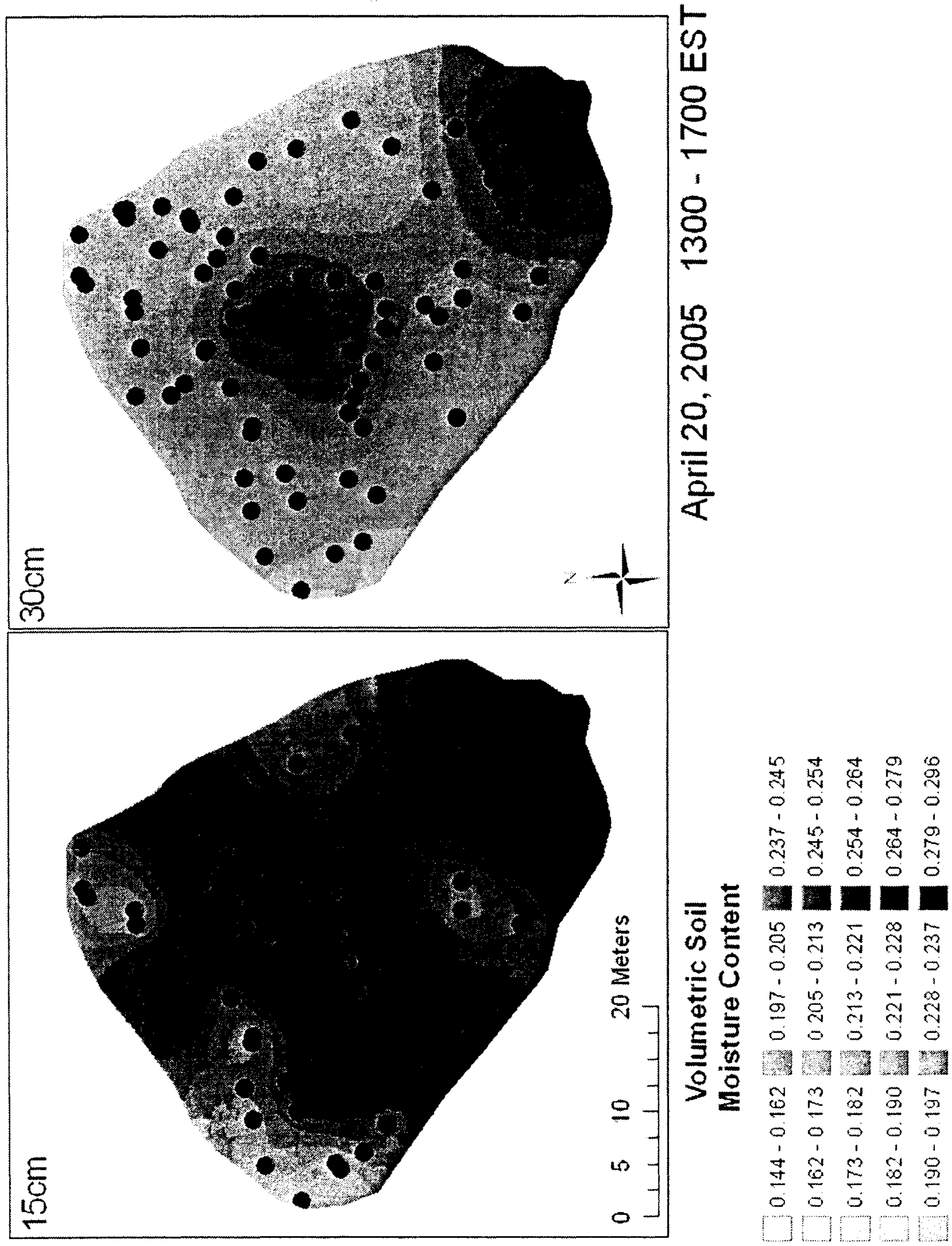
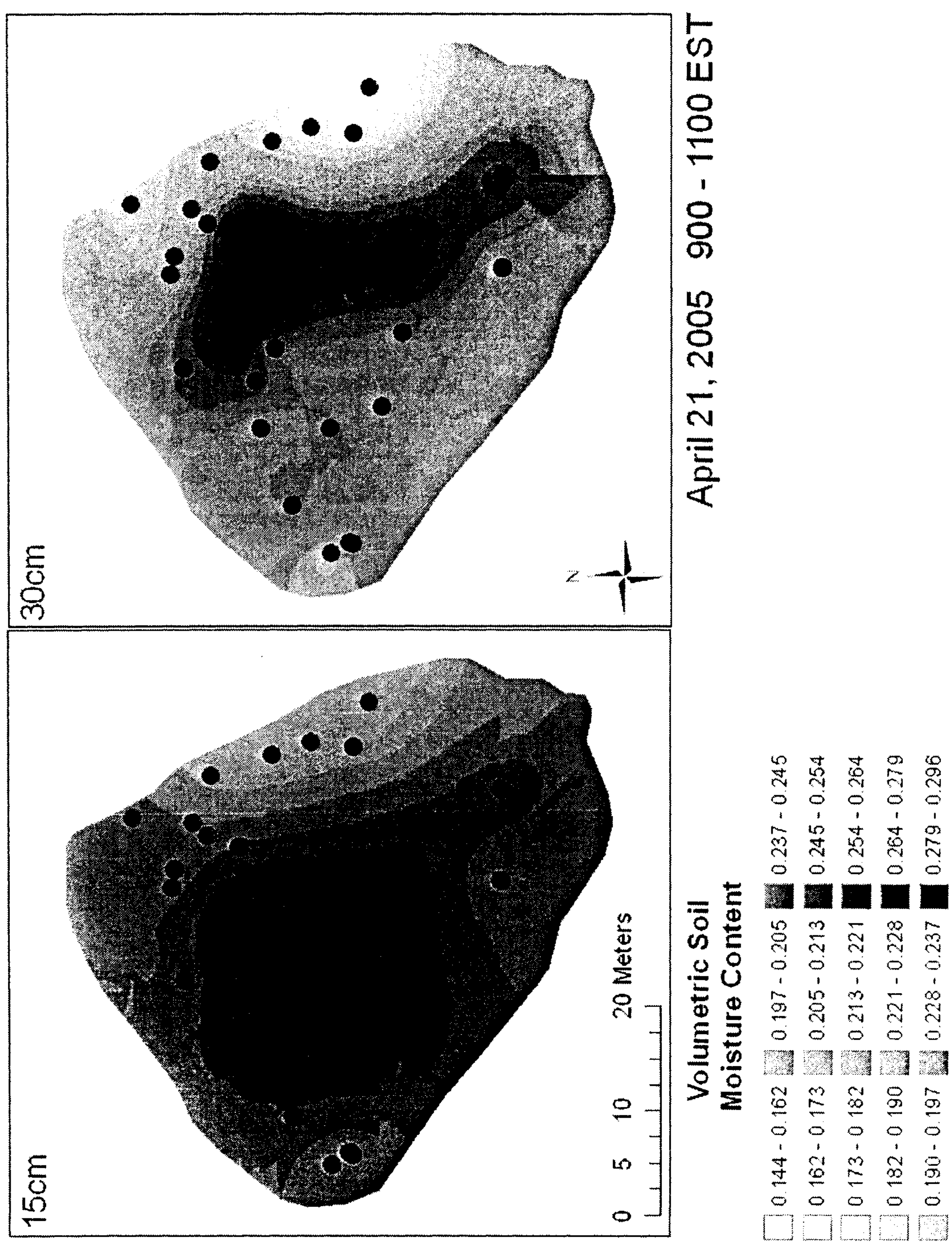
APPENDIX I

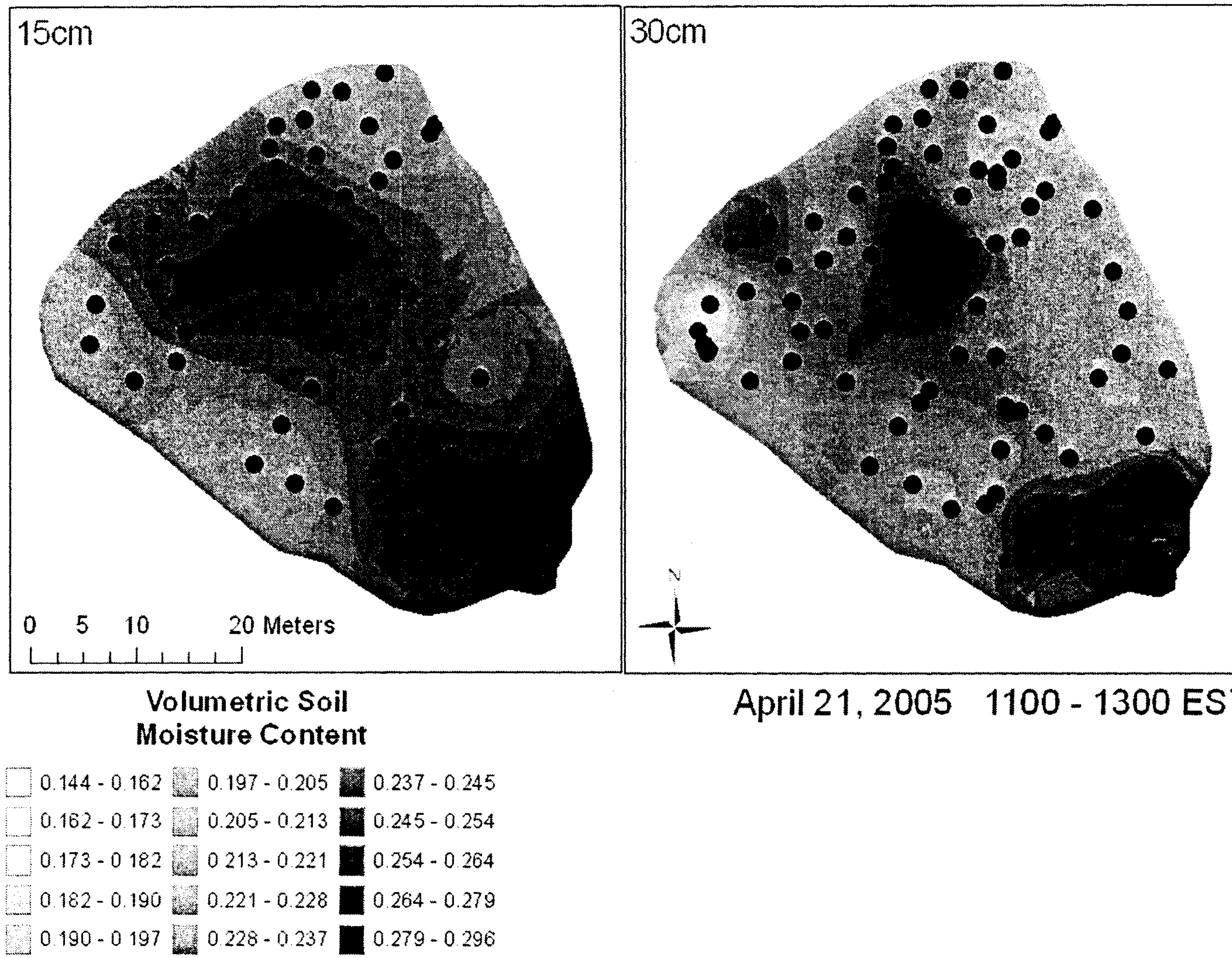
The evolution of soil moisture from April 20 – 23, 2005 at Glen Haffey



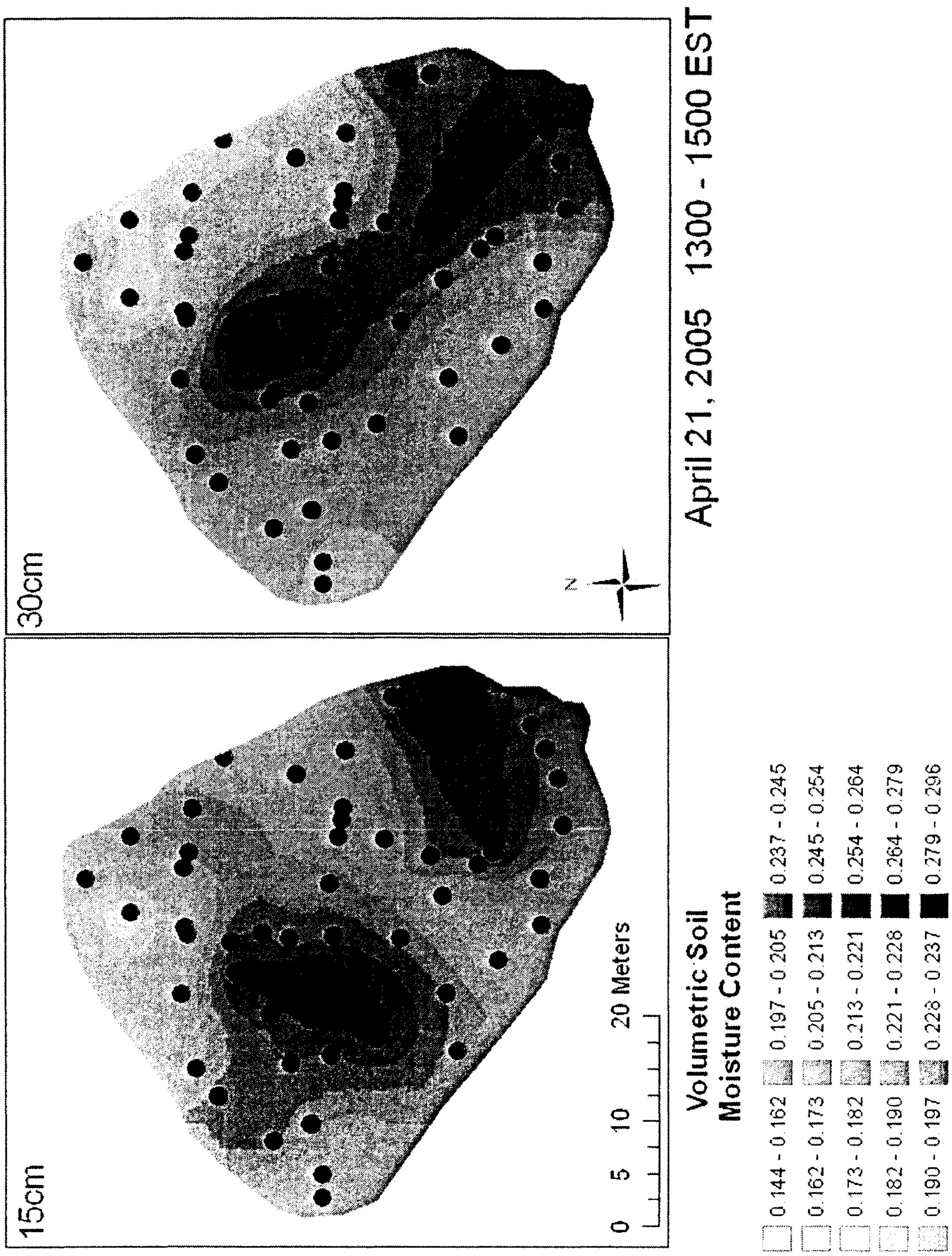

15cm
30cm
0 5 10 20 Meters
April 20, 2005 1100 - 1300 EST
Volumetric Soil Moisture Content
0.144 - 0.162
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296

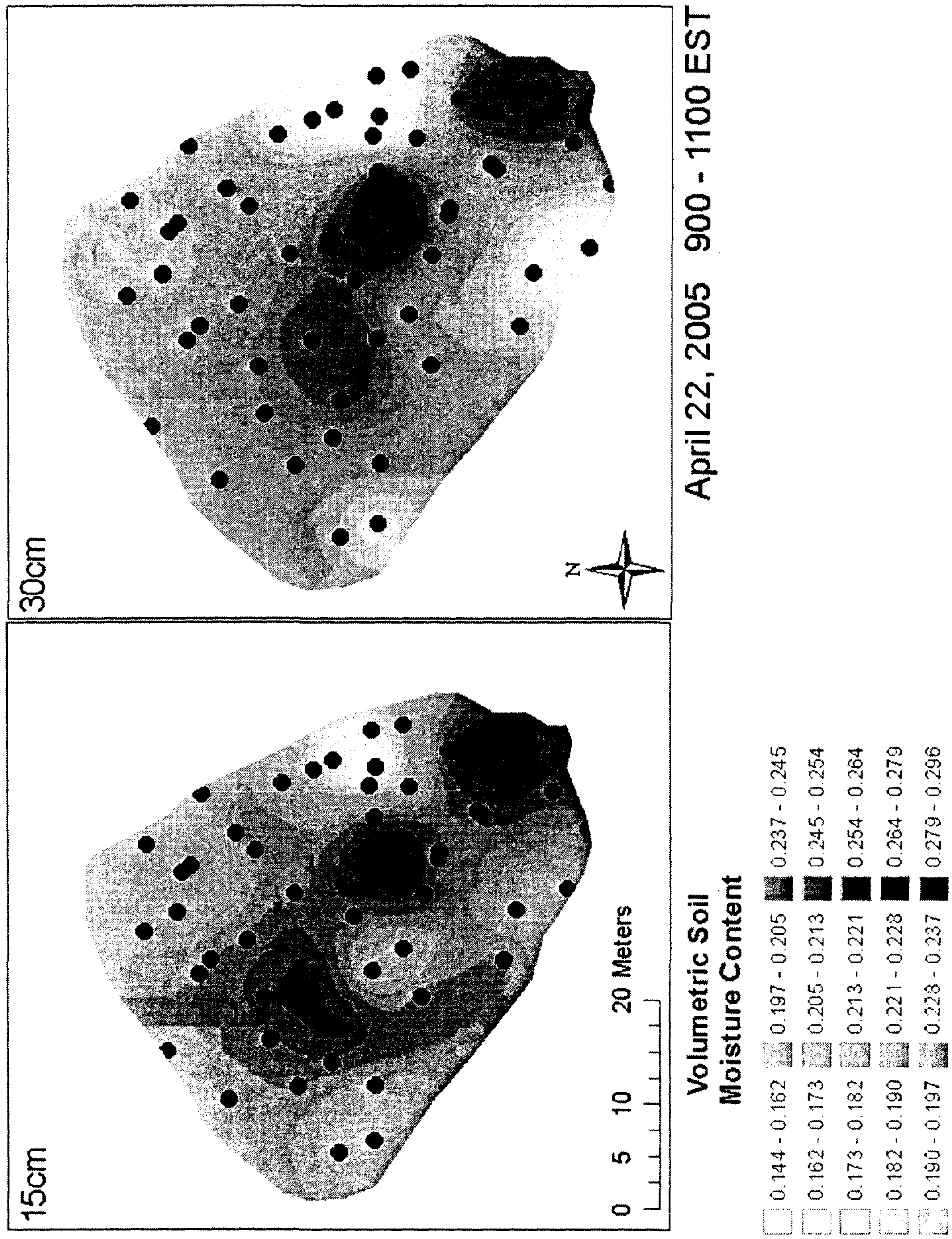

15cm
30cm
0 5 10 20 Meters
N
Volumetric Soil Moisture Content
0.144 - 0.162
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296
April 20, 2005 1300 - 1700 EST


15cm
30cm
0 5 10 20 Meters
N
Volumetric Soil Moisture Content
0.144 - 0.162
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296
April 21, 2005 900 - 1100 EST



April 21, 2005 1100 - 1300 EST


15cm
30cm
0 5 10 20 Meters
N
April 21, 2005 1300 - 1500 EST
Volumetric Soil Moisture Content
0.144 - 0.162
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296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30cm
0 5 10 20 Meters
N
April 22, 2005 900 - 1100 EST
Volumetric Soil Moisture Content
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0.162 - 0.173
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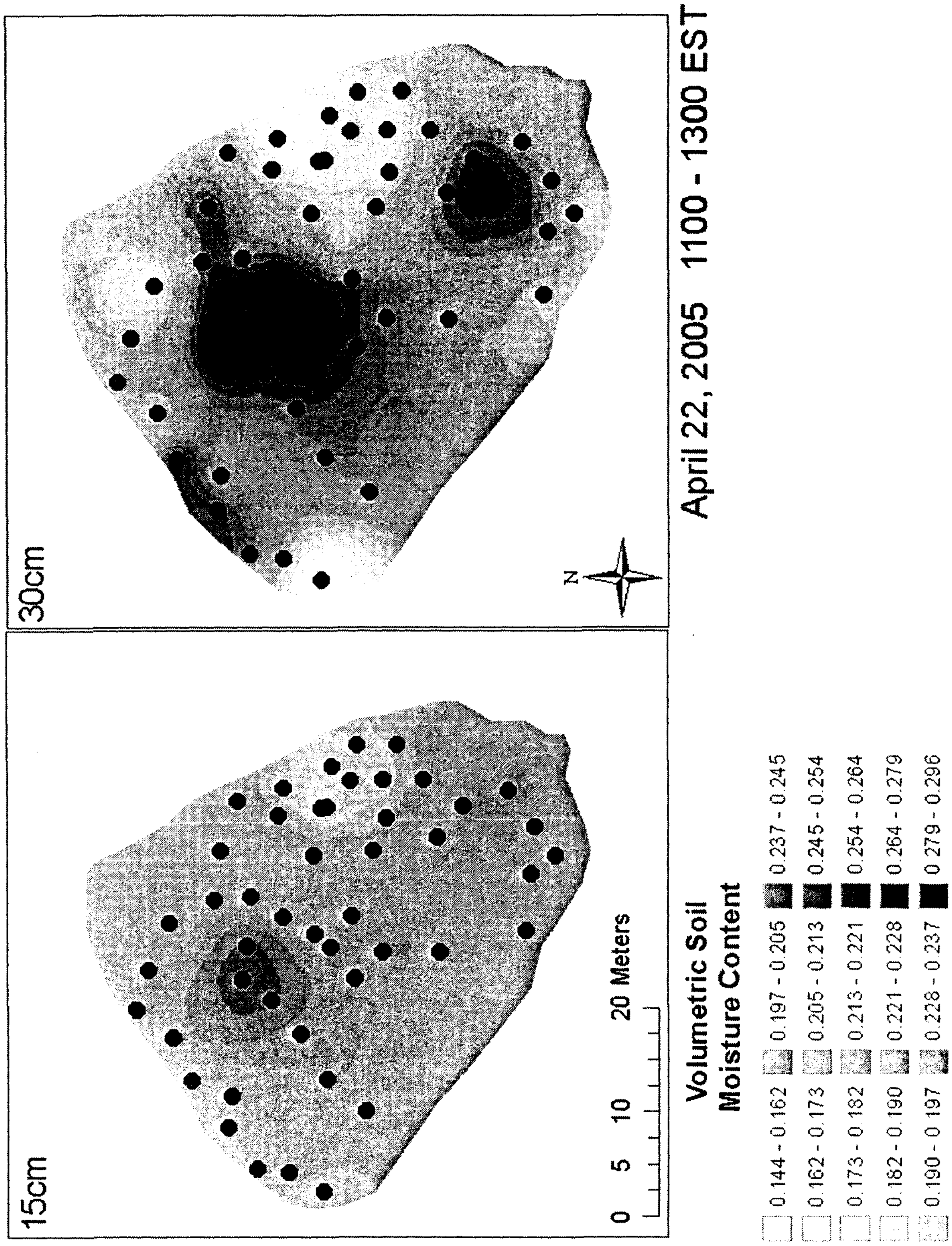
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296


15cm
30cm
0 5 10 20 Meters
N
April 22, 2005 1100 - 1300 EST
Volumetric Soil Moisture Content
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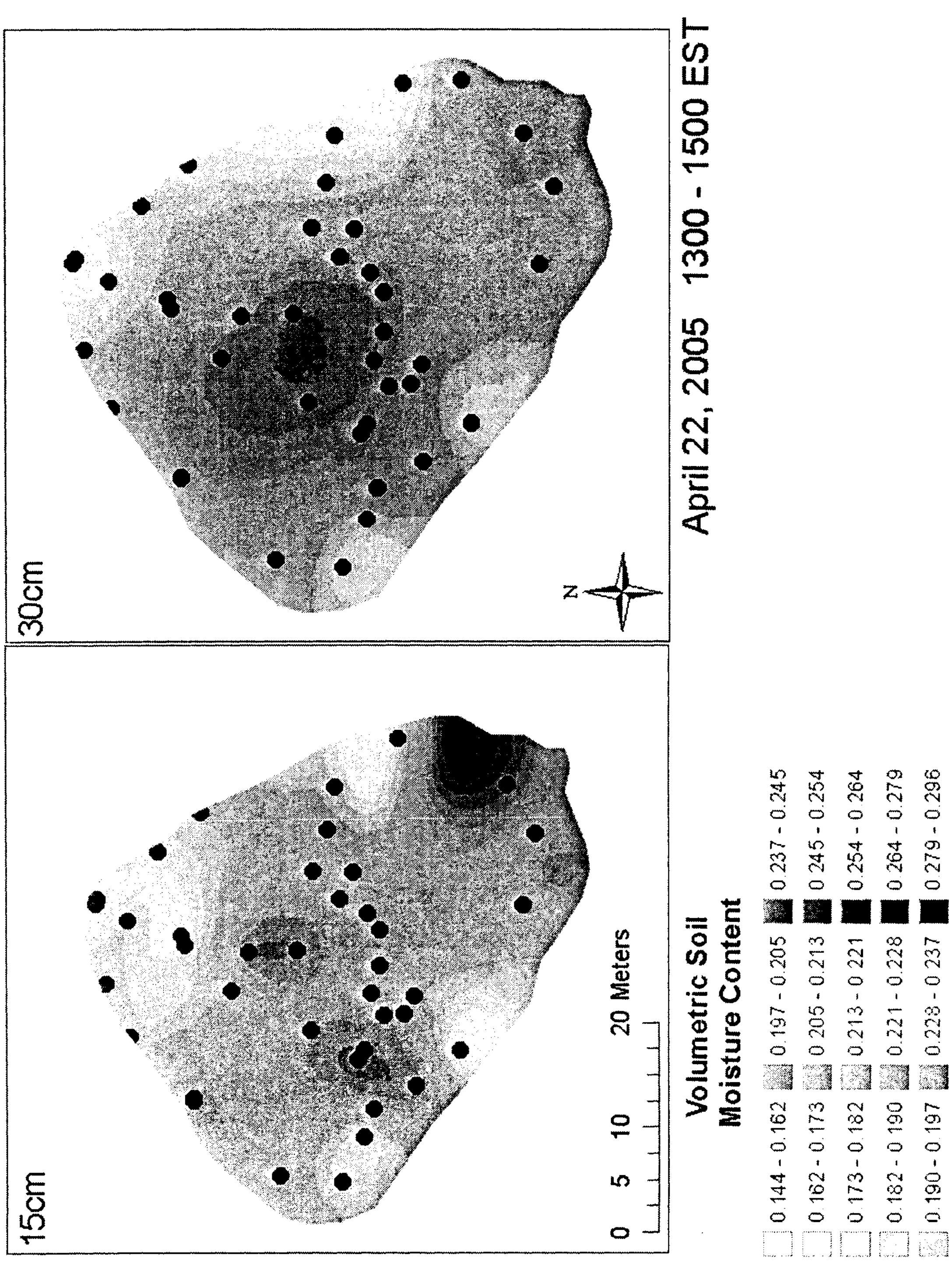
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296


15cm
30cm
0 5 10 20 Meters
N
April 22, 2005 1300 - 1500 EST
Volumetric Soil Moisture Content
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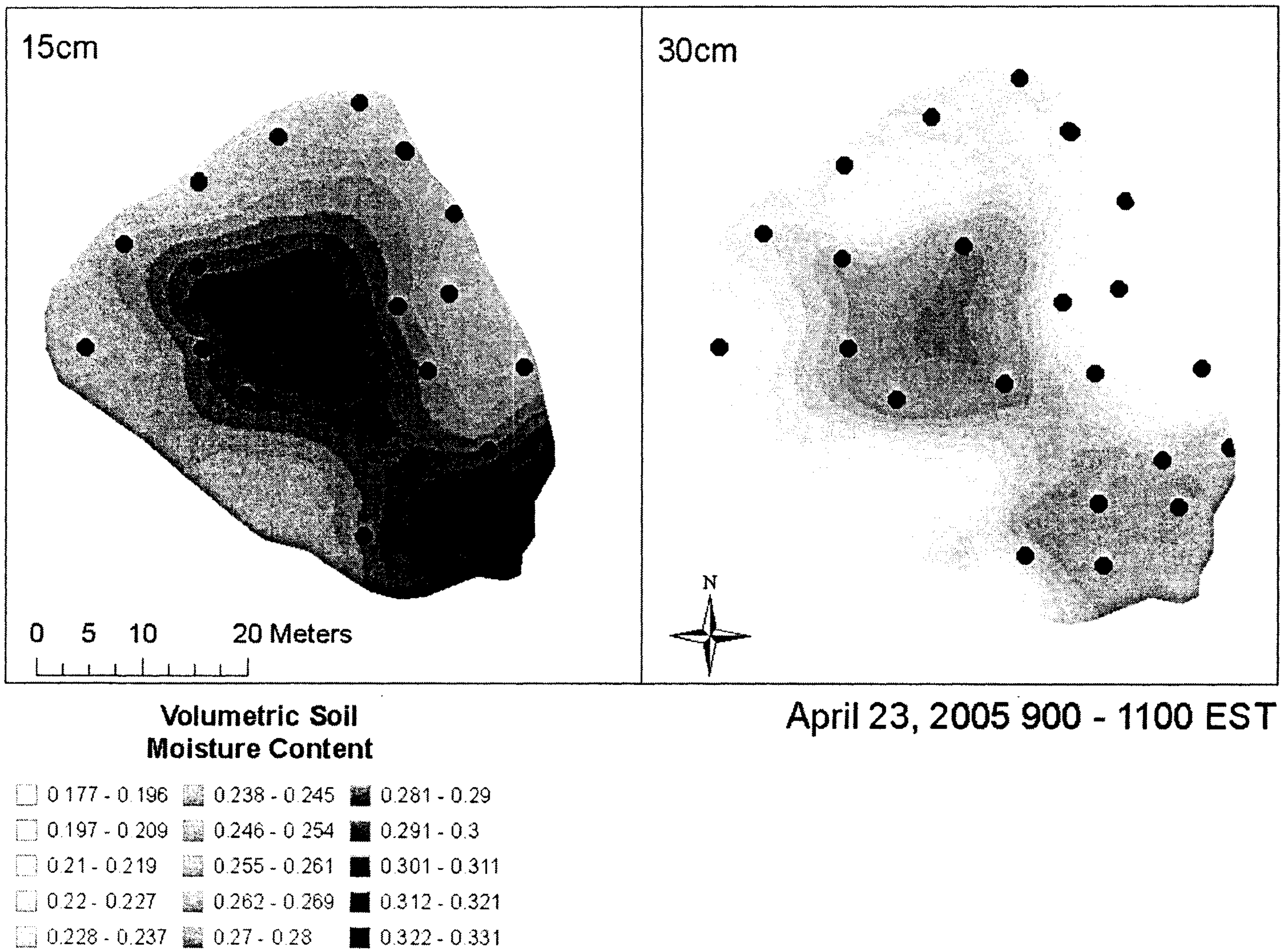
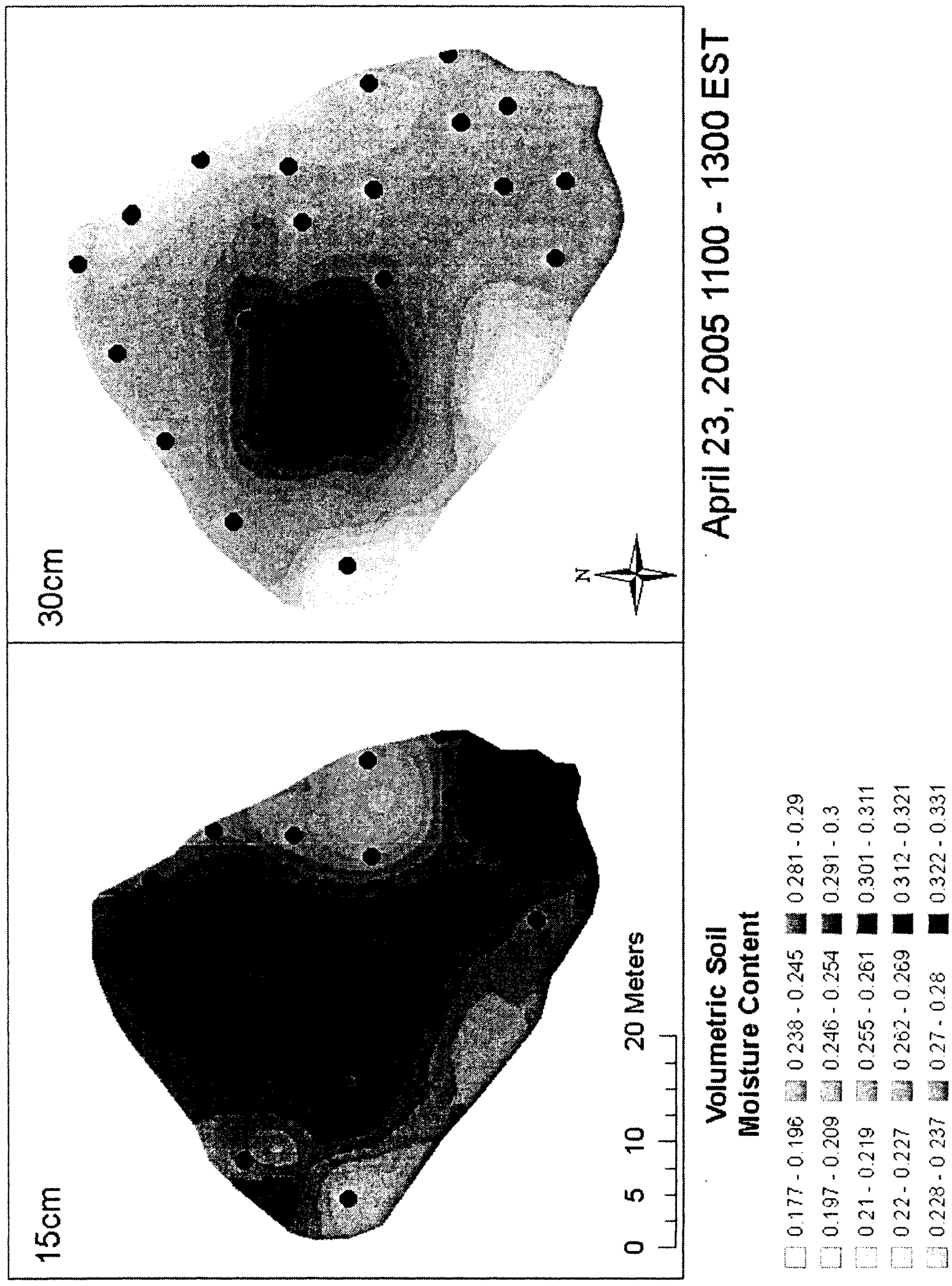
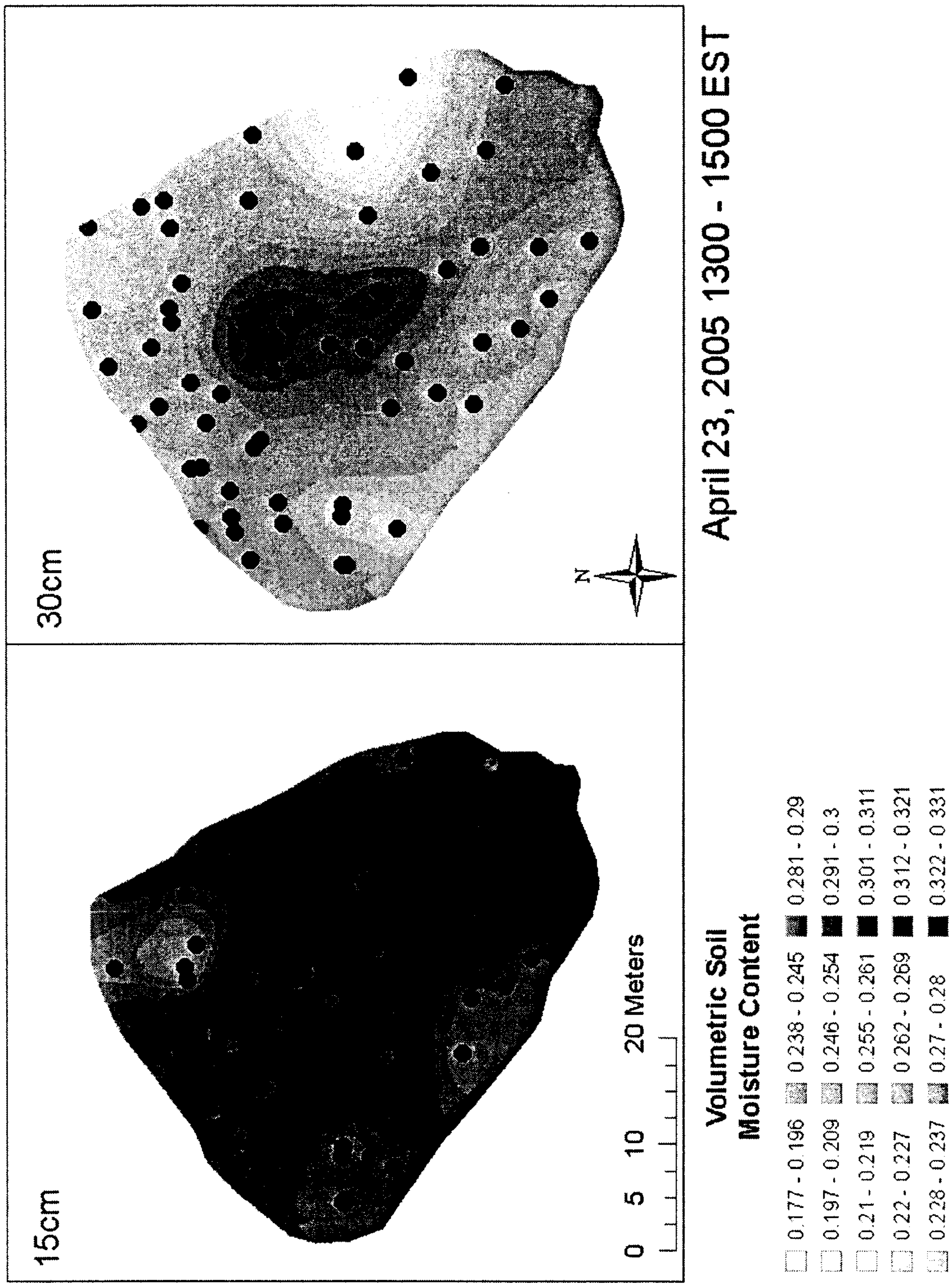
0.162 - 0.173
0.173 - 0.182
0.182 - 0.190
0.190 - 0.197
0.197 - 0.205
0.205 - 0.213
0.213 - 0.221
0.221 - 0.228
0.228 - 0.237
0.237 - 0.245
0.245 - 0.254
0.254 - 0.264
0.264 - 0.279
0.279 - 0.296


15cm
30cm
0 5 10 20 Meters
N
Volumetric Soil
Moisture Content
0.177 - 0.196
0.197 - 0.209
0.21 - 0.219
0.22 - 0.227
0.228 - 0.237
0.238 - 0.245
0.246 - 0.254
0.255 - 0.261
0.262 - 0.269
0.27 - 0.28
0.281 - 0.29
0.291 - 0.3
0.301 - 0.311
0.312 - 0.321
0.322 - 0.331
April 23, 2005 900 - 1100 EST


15cm
30cm
0 5 10 20 Meters
N
April 23, 2005 1100 - 1300 EST
Volumetric Soil Moisture Content
0.177 - 0.196
0.197 - 0.209
0.21 - 0.219
0.22 - 0.227
0.228 - 0.237
0.238 - 0.245
0.246 - 0.254
0.255 - 0.261
0.262 - 0.269
0.27 - 0.28
0.281 - 0.29
0.291 - 0.3
0.301 - 0.311
0.312 - 0.321
0.322 - 0.331


15cm
30cm
0 5 10 20 Meters
N
April 23, 2005 1300 - 1500 EST
Volumetric Soil Moisture Content
0.177 - 0.196
0.197 - 0.209
0.21 - 0.219
0.22 - 0.227
0.228 - 0.237
0.238 - 0.245
0.246 - 0.254
0.255 - 0.261
0.262 - 0.269
0.27 - 0.28
0.281 - 0.29
0.291 - 0.3
0.301 - 0.311
0.312 - 0.321
0.322 - 0.331

APPENDIX II

Sensitivity of Range length (m) to changes in number of lags and bin size

Number of Bins	10	11	12	13	14	15	16
Lag Size m)							
1	9.92	10.98	11.85	12.98	13.95	14.98	15.95
1.2	11.91	13.17	14.22	15.58	15.58	15.95	15.07
1.4	13.90	15.37	16.60	15.75	14.92	14.46	14.31
1.6	15.88	15.89	14.58	14.31	14.20	13.76	13.13
1.8	15.41	14.32	14.21	13.34	13.18	12.19	12.24
2	14.12	14.14	14.06	12.84	12.53	12.68	13.45
2.2	13.50	13.41	12.65	12.18	12.62	13.84	15.96
2.4	13.26	12.38	12.14	12.75	14.25	17.36	25.03
2.6	12.54	11.96	12.31	14.38	18.00	29.91	41.54

April 19, 30 cm

Number of Bins	10	11	12	13	14	15	16
Lag Size (m)							
1	9.92	10.98	11.85	12.98	13.95	14.98	15.95
1.2	11.91	13.17	14.22	15.58	15.58	16.40	16.55
1.4	13.90	15.37	16.60	16.69	16.26	16.40	18.04
1.6	15.88	16.99	16.10	16.48	17.80	17.66	16.65
1.8	16.28	15.64	16.57	17.65	17.25	15.34	14.84
2	15.36	17.09	18.25	15.94	15.31	14.45	13.52
2.2	16.51	17.19	16.03	14.76	13.84	13.36	13.23
2.4	17.32	15.34	14.87	13.45	13.34	12.98	13.29
2.6	15.73	14.27	13.57	13.14	12.90	13.23	13.80

April 22, 15 cm

Number of Bins	10	11	12	13	14	15	16
Lag Size (m)							
1	9.92	10.98	11.85	12.98	13.95	14.98	15.95
1.2	11.91	13.17	14.22	15.58	15.58	17.98	19.14
1.4	13.90	15.37	16.60	18.17	19.53	20.98	22.33
1.6	15.88	17.56	18.87	20.77	22.31	21.51	21.54
1.8	17.86	19.76	21.34	21.59	21.91	20.16	21.19
2	19.49	21.95	21.53	20.45	21.01	20.15	19.89
2.2	15.48	15.78	15.83	16.02	16.12	15.84	15.75
2.4	15.85	15.87	16.17	16.04	16.02	15.80	15.72
2.6	15.82	16.01	16.10	15.87	15.96	15.55	15.39

April 22, 30 cm

Number of Bins	10	11	12	13	14	15	16
Lag Size (m)							
1	9.92	10.98	11.85	12.98	13.95	14.98	15.95
1.2	11.91	13.17	14.22	15.58	15.58	17.98	19.14
1.4	13.90	15.37	16.60	18.17	19.53	18.66	18.43
1.6	15.88	17.56	18.97	19.01	18.61	18.54	18.47
1.8	17.86	19.56	18.56	18.47	18.55	18.08	18.03
2	19.80	18.30	18.59	18.20	18.12	17.80	17.52
2.2	18.31	18.38	18.33	17.98	17.81	17.35	17.33
2.4	18.38	18.26	18.25	17.68	17.45	17.25	17.10
2.6	18.32	18.12	17.92	17.51	17.42	17.22	17.31

Actual Data

-	Range	Bin Size	Number of Lags	RMS	RMSS
April 19 - 15cm	14.82	2.30	11	0.016	1.018
April 19 - 30cm	14.66	2.20	14	0.030	0.992
April 22 - 15cm	19.36	2.58	12	0.021	1.067
April 22 - 30cm	19.22	2.35	16	0.022	1.011

VITA AUCTORIS

Peter R. Andrew-McBride was born July 14, 1979 in Windsor, ON. He attended Riverside Secondary School (Windsor, ON) and graduated in 1998. In 2003, he received a B.Sc. from Queen's University and in 2006, a M.Sc. from the University of Windsor.